

Uppsala Universitet
Naturgeografiska institutionen
Avdelningen för Hydrologi

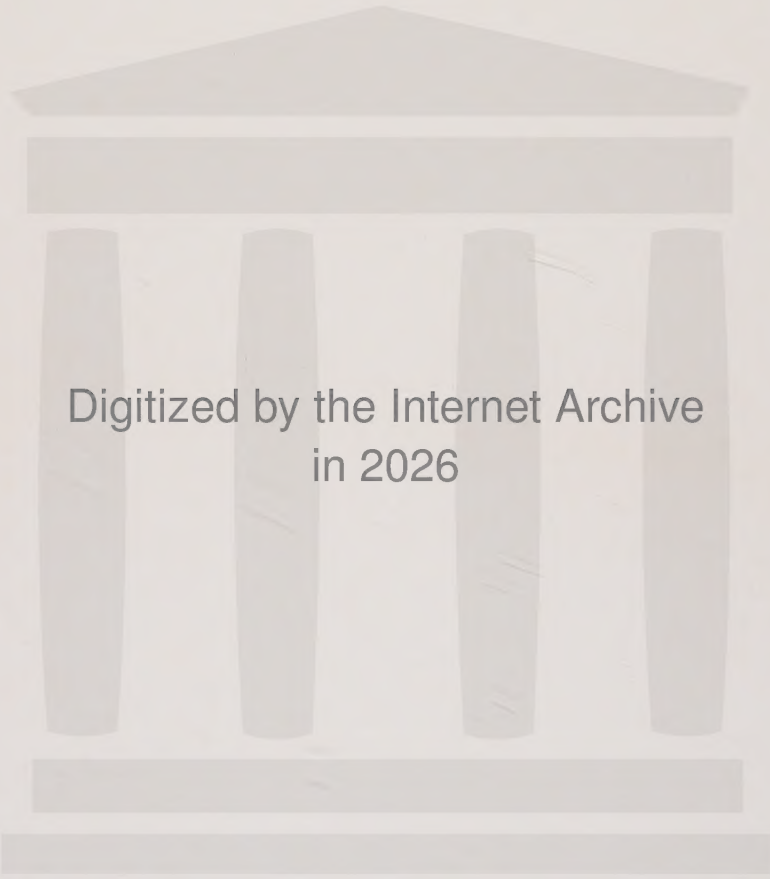


Allan Rodhe

The Origin of Streamwater Traced by Oxygen-18

Uppsala University
Department of Physical Geography
Division of Hydrology

Report Series A
No 41



Digitized by the Internet Archive
in 2026

Uppsala Universitet
Naturgeografiska institutionen
Avdelningen för Hydrologi



Allan Rodhe

The Origin of Streamwater Traced by Oxygen-18

Uppsala University
Department of Physical Geography
Division of Hydrology

Report Series A
No 41

Abstract

Rodhe, A., (1987): The origin of streamwater traced by oxygen-18. *Uppsala Univ., Dept. Phys. Geogr., Div. Hydrol., Report Series A 41*, 260 pp. Appendix 73 pp.

Along what flowpaths the rain- or meltwater flows before reaching the stream, and the related dynamics, are basic questions in many environmental problems. The stable isotope oxygen-18 was used as a tracer to separate stream discharge into flows originating from groundwater and rainwater or meltwater. Ten basins, with areas 0.03-6.6 km² and dominated by forested till soils, were investigated.

For 29 separated snowmelt runoff events, the volume fraction of groundwater, or pre-event water, ranged between 32% and 95% of the discharged volumes, with a median value of 61%. Of the total streamflow in spring, the fractions ranged between 41 and 86%. The groundwater fractions were generally larger during rainfall-generated events, with fractions always exceeding 68%. Groundwater fraction and maximum specific discharge were inversely related. Isotope enrichment in wet areas may have introduced an error when hydrograph separation of some events in basins of low relief was performed, as may also molecular exchange of oxygen-18 between overland-flowing water and groundwater.

The areal extent of saturated discharge areas was estimated assuming all new rainwater or meltwater in the streams to originate from the rainfall or melting on such areas. The discharge area of the basins was then found to range between 2% and 60%, median value 22%, for snowmelt events and between 0.2% and 17%, median value 3%, for rainfall events. From a few field surveys during snowmelt it was found that the estimations of the discharge areas were realistic, with the exception of the highest fraction.

Reservoir volumes for soil water and groundwater were estimated from the damping of input fluctuations in the oxygen-18 content during water flow through the basins. The calculated reservoir volumes of 200-300 mm correspond to mean transit times for water in the basins of 0.5-1.5 years.

Allan Rodhe, Division of Hydrology, Department of Physical Geography, Uppsala University, Västra Ågatan 24, S-752 20 Uppsala, Sweden.

ISSN 0281-8264

Uppsala 1987

© Allan Rodhe

Printed in Sweden by Fyris-Tryck AB, Uppsala 1987

CONTENTS

PREFACE	7
1 INTRODUCTION	9
1.1 Theories on streamflow generation	9
1.2 Outline of the present study	13
2 REVIEW OF LITERATURE ON STREAMFLOW GENERATION	16
2.1 Field measurements of water flows and potentials	19
2.1.1 Infiltration and overland flow	19
2.1.2 Subsurface flow	22
2.1.3 Soil water and groundwater potentials	25
2.1.4 Other experiments	28
2.2 Simulations with mathematical models based on Darcy's law ..	29
2.3 Interpretation of streamwater chemistry or isotope content	33
2.4 Conclusions from the review of literature	40
3 METHODS	42
3.1 Origin of streamwater traced by water chemistry or isotopes	42
3.2 Oxygen-18 as a tracer	48
3.2.1 General occurrence	48
3.2.2 $\delta^{18}\text{O}$ of precipitation related to the present study ..	52
4 SITE DESCRIPTIONS	56
4.1 Aspåsen	56
4.2 Buskbäcken	58
4.3 Gårdsjön	59
4.4 Nåsten	60
4.5 Stormyra	62
4.6 Svartberget	63
5 MEASUREMENTS	70
5.1 Field measurements	71
5.1.1 Precipitation	71
5.1.2 Snowpack	72
5.1.3 Groundwater	74
5.1.4 Streamwater	74
5.2 Laboratory measurements	75
6 FRACTIONATION PROCESSES FOR OXYGEN-18 WITHIN A BASIN	76
6.1 Models for fractionation due to evaporation and exchange with water vapour in the atmosphere	76
6.1.1 Models for the $\delta^{18}\text{O}$ of the net evaporate	76
6.1.2 Model for isotopic change in a well mixed water body	80

6.1.3	Sensitivity of the fractionation model to meteorological conditions	83
	The steady-state value	85
	The difference between the $\delta^{18}\text{O}$ of the water and of the evaporating vapour	86
6.1.4	Experimental results related to the present study ...	88
6.1.5	Limitations of the model for isotopic change	89
6.2	Estimates of enrichment during water flow through a basin ..	90
6.2.1	Interception	92
6.2.2	Infiltration	93
	Non-ponding infiltration	93
	Infiltration from surface reservoirs	95
6.2.3	Soil water	97
	Bare soil	97
	Transpiration	99
	Experimental results reported in the literature concerning enrichment of soil water	100
6.2.4	Groundwater	101
6.2.5	Surface waters in discharge areas	102
	Wet areas connected to the stream and the stream itself	103
	Stream channel flow and surface water reservoirs	105
6.2.6	Snow	108
	Relationship between snowpack and meltwater $\delta^{18}\text{O}$	110
	The isotopic composition of a snowpack	110
	Possible mechanisms of snowpack enrichment by the atmosphere	111
	Areal variability of snowpack $\delta^{18}\text{O}$	115
6.2.7	Conclusions on enrichment during water flow through a basin	115
7	ISOTOPE CONTENT OF THE WATER INPUT	118
7.1	Precipitation	118
7.1.1	Seasonal variation	118
7.1.2	Variation within storms	119
7.1.3	Areal variability	120
7.1.4	Correlation with air temperature	121
7.2	Snow	124
7.2.1	Mass and isotope balance of a snowpack	126
	Calculations of the net isotopic flux from the atmosphere	127
	Isotopic composition of meltwater from basin mean snowpack $\delta^{18}\text{O}$	130
	Vertical ^{18}O layering of the snowpacks	132
7.2.2	Measured $\delta^{18}\text{O}$ of meltwater	137
7.2.3	Snowpack $\delta^{18}\text{O}$ in forest and open sites	138
7.2.4	Evaporation estimate	139
7.2.5	Conclusion on meltwater $\delta^{18}\text{O}$	140

8	STREAMWATER. ESTIMATES OF GROUNDWATER FRACTIONS	141
8.1	Snowmelt events	142
8.1.1	Isotopic situation	148
8.1.2	Groundwater $\delta^{18}\text{O}$	149
	Observed $\delta^{18}\text{O}$ of groundwater	152
	Conclusions on groundwater $\delta^{18}\text{O}$	154
8.1.3	Modification of flow and tracer changes during flow in the channel network	155
	Relationship between flood wave celerity and particle velocity	156
	Damping and delay of fluctuations in tracer concentration	157
8.1.4	Variations of groundwater fractions within the events Daily variations in groundwater fractions of streamflow	159
	Diel fluctuations in groundwater fractions	162
	Conclusions on diel fluctuations	167
8.1.5	Groundwater fractions of total discharged volumes ...	168
8.1.6	Extent of discharge areas	168
	Field surveys of discharge areas	169
8.1.7	Conclusions and comments on snowmelt events basin by basin	173
8.2	Rainfall events	176
8.2.1	Intensively sampled events	179
	Methodological conclusions from the intensively sampled events	187
8.2.2	Events with "inexplicable" variations of streamwater $\delta^{18}\text{O}$	189
8.2.3	Common features of the separated rainfall-generated events	193
8.2.4	Groundwater fractions of total discharged volumes ...	194
8.2.5	Extent of discharge areas	195
8.2.6	Conclusions on rainfall events basin by basin	196
8.3	Estimates of reservoir volumes and transit times by ^{18}O	197
8.3.1	Short description of the model	198
8.3.2	Results	199
9	UNCERTAINTY DISCUSSION	203
9.1	Sensitivity of the model to uncertainty in the input variables	203
9.1.1	Groundwater fractions	203
9.1.2	Fractions of discharge area	205
9.2	Uncertainty in the input variables	205
9.2.1	Groundwater fractions	205
	Influence of precipitation during snowmelt	207
	Time lag between meltwater input and streamwater sampling	209
	Molecular exchange between overland flow and groundwater	209

Enrichment by the atmosphere	210
9.2.2 Fractions of discharge area	212
9.3 Uncertainty of the calculated groundwater fractions and fractions of discharge area	214
9.3.1 Groundwater fractions	214
9.3.2 Fractions of discharge area	215
9.4 Conclusions on uncertainty	216
10 CONCLUSION AND DISCUSSION ON STREAMFLOW GENERATION	218
10.1 Conclusions	218
10.1.1 Total volume fractions of groundwater	218
10.1.2 Extent of discharge areas	222
10.1.3 Reservoir volumes	224
10.1.4 Methodological problems	225
10.2 Discussion	227
10.2.1 Hypothesis for streamflow generation in till soils ..	228
10.2.2 Limitations of the estimates of discharge area by ^{18}O	229
10.2.3 Time variation of the groundwater fraction and the basin fraction of discharge area	230
10.2.4 Mechanisms for groundwater discharge response to water input	236
10.2.5 The origin of stormflow	240
10.2.6 Transit times for water in a hillslope	242
REFERENCES	247

APPENDICES IN A SEPARATE VOLUME:

APPENDIX 1	Figures showing $\delta^{18}\text{O}$ of precipitation and streamwater
APPENDIX 2	Snowmelt events
APPENDIX 3	Rainfall events
APPENDIX 4	Influence of precipitation on hydrograph separations during snowmelt
APPENDIX 5	Molecular exchange between overland flow and groundwater

PREFACE

The list of those who have actively contributed to this work is a long one. Erik Eriksson, now professor emeritus, was a great source of inspiration to me at the Division of Hydrology in the 1970s. With his broad experience and his ability to find imaginative interpretations to processes in nature, he made the subject of this thesis very exciting to me. I wish to thank him for his supervision and his always encouraging interest in the study. I also want to thank Professor Lars Bengtsson, who later took over the supervision, for his constructive and detailed criticism, which greatly improved the manuscript.

The study was financially supported by the Swedish National Sciences Research Council, NFR, in the form of a project in which I worked in cooperation with Rajinder Saxena, who helped me at all stages of the work.

The thesis is based on the work of the many helpers who carried out the water sampling in the basins: Rune Vedin in the Aspåsen basin, Birgit Persson, Anders Bergstad and Urban Ahlström in Buskbäcken, Katarina Hedlöf and the field staff of the Lake Gårdsjön project in the Gårdsjön basins, Monia and Per Eriksson in Nåsten, Eva Springer and Helga Henriksson in Stormyra, and Elon Manfredsson and the staff of Vindeln Forest Research Station in Svartberget. Thanks for all the bottles of water! Also thanks to Torbjörn Nilsson, Göran Lindström and Kevin Bishop, for various water sampling expeditions.

The project was made possible by my having access to the Mass Spectrometry Laboratory at the Division of Hydrology, under the leadership of Dr Jan Olof Burgman. I wish to express my gratitude to him and his staff.

I also want to thank all those who have been employed in the project during various periods and assisted me with field and laboratory measurements and in the initial processing of data: Maigret Bengtsson, Barbro Johansson, Gunnel Alvenäs, Judith Ferenczi, and Anders Ekstav. Thanks also to my other colleagues for fruitful discussions and to Harald Grip, Göran Nyberg and my brothers Henning and Lasse for reading and criticising parts of the manuscript.

Much of my knowledge about this subject has grown out of the discussions with students in the lecture hall and in the stimulating cooperation with students during many years' field courses. I also want to thank the students for collecting some of the data used in this study.

The figures were drawn by Kjerstin Andersson, Margareta Andersson, and Richard Heikenskjöld, and the photographic work was done by Assar Lindberg. Thanks to them and to the many other persons at the Department of Physical Geography who have assisted me in the work, and also to Donald MacQueen and Don Pierson who corrected my English.

I am especially grateful to Kristina Nilsson who not only typed and skilfully edited the manuscript, but also discovered several obscurities in the text and gave valuable linguistic advice.

Finally my thanks go to my wife Hedda, who has always been interested in this work and assisted me in various ways. Thanks for all support and for the many nice water sampling picnics we have had together over the years, accompanied by a growing number of children.

Uppsala 6 May 1987

Allan Rodhe

1 INTRODUCTION

The mechanism by which rainfall or snowmelt in a basin is transformed into streamflow, i.e., the runoff process or the process of streamflow generation, is not yet fully understood. Three hundred years after the establishment of the mass balance for water in a basin (generally attributed to Perrault 1674) many questions remain concerning the water flow from rainfall to the stream. The basic questions which need to be answered, and related to climate and physiography, are: What paths do the water particles take through the basin, what are the magnitudes of the flows and residence times of the water on the surface and in different soil layers, and how can these processes be formulated mathematically? A better knowledge of the runoff process is urgently needed for a further development of mathematical runoff models and, particularly, for an understanding of the chemical changes taking place in the water during its flow through the basin to the draining stream.

In this study, one aspect of the runoff process is investigated: the origin of streamwater. Estimates are made of the contribution to streamwater flow by fresh rainwater or meltwater, on the one hand, and by groundwater, on the other. The estimates are made using the environmental stable isotope oxygen-18 (^{18}O) as a tracer.

1.1 Theories on streamflow generation

What is considered today to be the traditional picture of streamflow generation is the infiltration theory formulated by Robert E. Horton during the early decades of this century (Horton 1933, 1945). The theory was advocated as a tool for making better forecasts of streamflow from rainfall data than the existing theories, which viewed runoff either as a certain fraction of rainfall or as the difference between rainfall and evapotranspiration. These latter approaches could at best give information about long-term mean flow, not about individual runoff events (Horton 1933). Horton introduced the term "rainfall excess" as the amount of rainfall which falls at intensities exceeding the infiltration capacity of the soil. He thereby

stressed the role of soil properties in determining the infiltration capacity, and rainfall intensity as giving more information than just rainfall amount.

The rainfall excess is assumed to generate overland flow, some of which is captured as surface detention while the remainder flows to the draining stream. Stormflow in streams is considered to originate from overland flow from the whole basin. This overland flow is added to outflowing groundwater which, depending on soil moisture conditions, is in varying degrees affected by the rainfall that caused the stormflow. Between the stormflow events the streams are fed by groundwater. From the storm event hydrograph and the recorded rainfall the infiltration capacity of the basin can be calculated and used later in forecasting streamflow for given rainfalls.

This view of the runoff process has formed the basis for applied hydrology up to the present day. The widely used unit hydrograph (Sherman 1932) is an early forecasting model based on the Hortonian infiltration theory. Alternative views, stressing the dynamic nature of subsurface flow, existed early (Hoyt 1942, Hoover and Hursh 1943) but seem to have had little influence on hydrological practice. However, during the 1960s and 1970s the validity of the Hortonian infiltration theory in humid temperate areas was questioned, even though it still forms a common visualization of streamflow generation.

In the early 1960s, researchers in the eastern United States suggested that rainfall over only a minor part of the drainage basin contributed to stormflow. Starting from Hortonian theory, Betson (1964) developed an infiltration equation and interpreted the results such that rainfall intensity exceeded infiltration capacity over only a fraction of the basin, which he called "partial areas". These were supposed to be mainly those regions close to the stream where the infiltration capacity is low, or zero, due to high soil moisture content. On the basis of hydrograph analysis the Tennessee Valley Authority (1965) developed a time distribution function for storm runoff from precipitation. The best fit to observed stream discharge was obtained when runoff was assumed to originate from rainfall over "variable runoff-producing areas", i.e., areas whose extension varies with time due to changes in soil moisture status. The runoff process is thoroughly discussed by Hewlett and Hibbert (1967) on the basis of their concept of "variable source areas" (introduced by Hewlett 1961). The limitation of the Hortonian infiltration theory was demonstrated in a later

study by Hewlett et al. (1977). Regression analyses of a 30-year series of recorded precipitation showed that storm runoff was strongly correlated to rainfall volume and that rainfall intensity had little influence upon stormflow production.

In a classic field study of runoff production Dunne and Black (1970a, b) found that overland flow occurred only on areas where the water table had risen to the soil surface. They introduced the term "saturation overland flow" to distinguish this overland flow from that described by Horton, which occurs on unsaturated soil (although the very topmost layer must be saturated for any type of overland flow to occur). Dunne and Black stressed that the areal coverage of the partial areas depends on the groundwater level, which thus determines the hydrological response of a basin to a certain rainfall. According to them, stormflow in streams is dominated by saturation overland flow, mainly caused by rainfall on saturated areas. The role of groundwater in streamflow generation is passive. It regulates the extension of the saturated areas, which in turn determines the occurrence of overland flow and thus streamflow production. This view is strongly supported by Freeze (1972a, b; 1974) who argued from results of simulations with physically based mathematical models for hillslope runoff.

The partial area concept discussed by Hewlett and Hibbert (1967) is different. They emphasized the active role of soil water and groundwater in streamflow generation. Their "variable source areas" constitute saturated or near-saturated areas connected to the stream. These areas transmit unsaturated subsurface flow from the upper parts of the hillslopes to the stream. The subsurface flow is considered to be an important component of stormflow, and a large fraction of the streamwater discharged during a storm is assumed to consist of old water, transported in the soil by so-called "translatary flow". The existence of this subsurface stormflow has been under considerable debate (Freeze 1974, Knisel 1973, Hewlett 1974, Martinec 1975, Ward 1984).

Several studies using environmental isotopes have shown that fresh rainwater generally constitutes a minor part of the streamwater during runoff events (e.g. Sklash and Farvolden 1979, see Chapter 3). These studies give little information on the flow processes within the basin, but they show unequivocally the existence of some sort of "subsurface stormflow". However, many recent papers, discussing the runoff process, make no reference to isotope studies (see for instance the two

textbooks "Hillslope Hydrology" (Kirkby 1978) and "Hydrological Forecasting" (Anderson and Burt 1985), both of which contain several chapters on streamflow generation).

The concept of "variable source area" is somewhat similar to the concept of "variable discharge area for groundwater", which has formed the basis for a great part of the hydrological research in Sweden since the late 1960s. Noting that the water table in Swedish till soils largely follows the topography, Gustafsson (1968), using an earlier mathematical model of groundwater flow (Gustafsson 1946), showed that a basin can be divided into recharge and discharge areas for groundwater, depending on whether the flow is directed into or out of the groundwater zone. The lower part of certain slopes, certain low-lying areas adjacent to the stream, and the stream itself, constitute discharge areas. The upper parts of slopes and elevated areas in general are recharge areas. The upward and downward directions of groundwater, caused by an undulating water table, was shown theoretically by Hubbert (1940) and given some field evidence from piezometer measurements by Kirkham (1947). Further support comes from the mathematical analysis of groundwater flow by Tóth (1963), who identified different groundwater flow systems in the vertical plane. Gustafsson (1968) discussed field characteristics of the discharge areas, such as vegetation, soil profile, and agricultural practice.

The view of recharge and discharge areas, based on steady-state flow conditions, has been developed into a general picture of streamflow generation in Swedish till soils (e.g. Eriksson, 1985). Stormflow in the streams is mainly caused by outflowing groundwater from discharge areas due to infiltration in recharge areas. The other component of stormflow is the rainfall on the discharge areas, flowing as overland flow to the stream together with the out-flowing groundwater. This view, which is discussed in some detail in the last chapter of this report, is the hypothesis for the present study.

Recent literature reviews on the process of streamflow generation are those by Dunne (1983) and Ward (1984) and, for tropical areas, the one by Walsh (1980). The above-sketched hypothesis for streamflow generation in Swedish till soils is discussed in detail by Grip and Rodhe (1985), who also relate the flow pattern to chemical processes in the water.

To conclude this presentation of the growth of theories for streamflow generation, one can state that the general view

among researchers today is that the Hortonian model is of little relevance in humid temperate areas and that some sort of areally variable active area has to be considered. There are, however, contradicting opinions in the international literature concerning the processes involved, particularly concerning the role of soil water and groundwater flow in generating stream stormflow. This disagreement is certainly partly due to different topographical, geological and climatological conditions in the areas where field studies have been undertaken, and to the problems of generalizing the results with regard to these factors. But it may also reflect the difficulties in interpreting field data so as to understand the processes involved in one single investigated basin.

1.2 Outline of the present study

The aim of the present study is to contribute to the understanding of the processes of runoff by investigating the roles of new rain- or meltwater and groundwater in stream runoff. Streamflow in ten small basins and sub-basins is divided into these two flow components using environmental ^{18}O as a tracer. The separation of streamflow into components is commonly called hydrograph separation (e.g. Linsley et al. 1949). Here isotopes of the water are used for the separation, hence isotopic hydrograph separation.

Since stable isotopes have seen little use in hydrological studies in Sweden, their application to problems concerning the process of runoff is investigated in some detail. Particular emphasis is placed on the methodological problems associated with isotopic hydrograph separation and the problems involved in using isotopes to model the flow of water and chemical compounds through a basin.

The thesis is arranged as follows:

Chapter 2 is a literature review of studies on the process of runoff. The review is organized according to the methods of investigation. Field studies on the various stages of the process are treated in rather great detail. Then follows a discussion on the more or less integrated results of the various processes involved, obtained by mathematical modelling of hill-slope flow and by isotope studies.

Chapter 3 presents the method used for isotopic hydrograph separation and for interpreting the estimated contributions of new rain- or meltwater in terms of hydrologic conditions within the basin. An introductory presentation of oxygen-18 as a tracer in hydrology is also given.

The investigated basins and the measurements are described in Chapters 4 and 5 respectively.

Chapter 6 is a discussion on fractionation processes for ^{18}O during water flow through a basin. The intention is to identify, and if possible quantify, isotopic changes which may affect the interpretation of the ^{18}O -content of water in terms of hydrological processes. A model for isotopic changes of water exposed to the atmosphere is developed and its sensitivity to variations in parameters and input data is discussed. The rate of isotopic change under various conditions encountered by the water (as liquid water or as snow) is judged from experimental results reported in the literature and from applications of the model. In these theoretical considerations, climatic, hydrologic and physiographic conditions are chosen so as to roughly represent the basins in the present study.

As a background to the hydrograph separations by ^{18}O , and to the use of ^{18}O for hydrological and hydrochemical modelling, observed data on the ^{18}O -content of precipitation and snowpack are presented in Chapter 7. The data on precipitation are discussed briefly in order to give an idea of the temporal and areal variability related to the present study. The snowpack data are analysed in more detail in an attempt to make generalizations on the relationship between the ^{18}O -content of meltwater, i.e., of the water input to the soil, and the ^{18}O -content of winter precipitation.

The hydrograph separations by ^{18}O are presented in Chapter 8. Here, runoff events generated by snowmelt and by rainfall are treated separately. Many methodological problems are discussed, such as changes of the ^{18}O -content of the groundwater component of streamflow during runoff events, the time lag between the changes in tracer concentration and discharge during streamflow, and damping in the channel network of short-term fluctuations in tracer concentrations. The extent of discharge areas as estimated from ^{18}O is compared with the results of a few field surveys. A simple model for estimates of reservoir volumes and transit times for water in a basin is also presented and applied. Contrasting with the hydrograph separations, which are

event studies, these estimates are based on the long-term variations (several months) of the ^{18}O -content of precipitation and streamwater. The discussion on the hydrograph separations in this chapter is mainly restricted to methodological questions. Conclusions concerning the process of runoff are drawn in Chapter 10.

The sensitivity of the isotopic hydrograph separations to errors in the input variables is discussed in Chapter 9. Attempts are made to quantify random and systematic errors in the input variables and to give figures on the accuracy of the results obtained.

In Chapter 10, finally, the results of the hydrograph separations are summarized and related to the hydrologic characteristics of the events. The mechanism by which rain- or meltwater generates streamflow and the role of different parts of a basin in this process is discussed on the basis of a hypothesis given for the runoff process in forested till soil.

An appendix is printed as a separate volume. It contains diagrams on the temporal variation of the isotope content of precipitation and streamwater during the investigated periods, diagrams of all separated snowmelt events, and annotated diagrams of the rainfall events not presented in the main text. It further contains two appendices to the uncertainty discussion in Chapter 9. In one of them, possible systematic errors due to tracer changes by molecular exchange between overland-flowing water and groundwater are discussed rather thoroughly.

2 REVIEW OF LITERATURE ON STREAMFLOW GENERATION

When analysing field results on the process of runoff, the first step is, of course, to interpret the results in terms of processes within the investigated domain, which may be a plot, a hillslope section or a basin. The difficulties may be very great even at this stage, and erroneous conclusions are certainly drawn in many cases. When it comes to generalizations of the results for applications at other sites, the difficulties might seem insurmountable, considering the complex structure of a natural basin and the large variability in topography, geology and climate between different basins. But it is necessary to make generalizations, and since the physical laws governing water storage and flow are universal, it is possible. At the present stage of knowledge there exist a few very general views on the process of runoff, commented upon in Chapter 1. These views are useful when one is to interpret field results or predict general flow patterns for water. But when one tries to understand the details of the runoff process in a particular basin, one has to consult the sum of experience from various investigations. With this in mind a review of studies that enlighten the process of streamflow generation is given in this chapter. The various results are presented in rather great detail, and general conclusions are drawn at the end of the chapter. An important question when making the review has been to find documented field evidence of subsurface stormflow and explanations of the mechanism governing such flow. The review is restricted almost exclusively to literature in English, mostly reporting investigations in humid, temperate areas in North America, Great Britain and Central Europe.

The hydrological conditions in most of the areas in which investigations have been reported differ from those in the Swedish till soils on crystalline rocks. One important difference is the shallow groundwater encountered in such Swedish areas, even in elevated localities. In many of the reported studies the groundwater table is close to the soil surface only in the valley bottoms, whereas shallow groundwater in the hillslopes is absent or develops only during wet conditions.

Knowledge of streamflow generation has been obtained by investigations using many different methods. The most important have been:

Measurement of infiltration capacity by infiltrometers
 Overland flow measurements from runoff plots
 Subsurface flow measurements by water collectors in pits or natural seepage faces
 Subsurface flow measurements using artificial tracers
 Detailed monitoring of soil water and groundwater conditions including mapping of surface saturated areas.
 Comparisons between rainfall and streamflow characteristics
 Mathematical modelling of soil water and groundwater flow in hillslopes
 Mathematical analysis of groundwater levels and flowpatterns on a basin-wide scale
 Hydrograph separation using water chemistry or environmental isotopes

The review is organized according to the above list. Special attention is given to results applicable to Swedish conditions and to the method used in the present study, i.e., hydrograph separation by environmental isotopes.

Some central terms used in the literature on the process of run-off are listed below. The terms are not always clearly defined, or they are defined in different ways by different authors. The definitions given here are those used in the present text.

Soil water. Water in the soil or bedrock having a pressure below that of the atmosphere. Occurs in the unsaturated zone.

Groundwater. Water in the soil or bedrock having a pressure above or equal to that of the atmosphere. Occurs in the saturated zone. Saturated zones may develop above impeding layers, underlain by an unsaturated zone. If such a saturated zone is permanent, deep, and of large areal extent, its water may be considered as groundwater (so-called perched groundwater). If a saturated zone overlying an unsaturated zone is very shallow and small, and if it occurs only occasionally during wet periods, the water is not considered to be groundwater.

Saturated area. Area in which the groundwater zone reaches the ground surface.

Recharge area. An area in which the flow is directed into the groundwater zone. (An area with downward flow in the uppermost groundwater.)

Discharge area. An area in which the flow is directed out of the groundwater zone. (Normally an area with upward flow in the uppermost groundwater).

Unsaturated flow. Flow of water having a pressure below that of the atmosphere.

Saturated flow. Flow of water having a pressure above or equal to that of the atmosphere. Saturated flow in the soil may occur in zones not considered as groundwater zones (see definition of groundwater).

Subsurface flow. Unsaturated or saturated flow in the soil or bedrock with flow paths directed towards the discharge area.

Subsurface stormflow. Subsurface flow that contributes significantly to the stormflow hydrograph in the stream.

Overland flow. Flow of water on the ground.

Saturation overland flow. Overland flow in saturated areas.

Hortonian overland flow. Overland flow in unsaturated areas, occurring when the rain or snowmelt intensity exceeds the infiltration capacity of the soil.

Runoff event. The increase of streamflow directly associated with a certain period of rainfall or snowmelt and the subsequent recession to about the initial value. The term is also used to denote the time period during which the above flow takes place.

Event water. Liquid water entering the basin during the event, i.e., new rain- or meltwater.

Pre-event water. Liquid water existing in the basin prior to the event.

Groundwater flow in a stream. Flow of water that has entered the stream after having passed the groundwater zone. In general discussions the term is used as a synonym to pre-event water,

although some of the groundwater in the stream may be new rain- or meltwater.

Flow of rain- or meltwater in a stream. Rain- or meltwater that has reached the stream as overland flow or channel interception. In general discussions the term is used as a synonym to event water, although some of the event water in the stream may have passed the groundwater zone and is thus a part of the groundwater flow in the stream.

Contribute to a runoff event. A sub-area in a basin contributes to a runoff event in the stream if rainfall or snowmelt on the sub-area in some way causes an increase of streamflow, either directly via water flow or indirectly via pressure propagation through unsaturated and saturated zones.

2.1 Field measurements of water flows and potentials

2.1.1 Infiltration and overland flow

There are numerous measurements of the infiltration capacity of soils reported in the literature, but few of them are made on a large-scale basis allowing conclusions to be drawn for a whole basin. An extensive study of infiltration capacities was performed in south-western England by Hills (1971) using cylindric infiltrometers. From comparisons with rainfall statistics he concluded that only a small fraction of the rainfall events (10-30%) can produce overland flow, and then only on bare or lightly vegetated compacted ground. A similar study, but aimed at covering one particular basin, was performed in northern England by Tricker (1981). Tricker gives a map of infiltration capacities in the 3.6 km² basin, according to which the soil in about 70% of the basin has an infiltration capacity greater than 20 mm h⁻¹, but no relation to rainfall rates is given.

Sharma et al. (1980) investigated the spatial variability of infiltration parameters in a 0.1 km² grassed basin of low relief in the central United States. Ring infiltrometer tests at 26 points gave infiltration capacities ranging from 4 to 130 mm h⁻¹, with a median value around 40 mm h⁻¹. The existence of a seasonal variation of the infiltration capacities in this basin was emphasized by Luxmoore (1983), who suggested that the infiltration capacity was considerably smaller in spring than during summer, when Sharma et al. performed their measurements. Development of cracks due to shrinkage of the soil mass was

considered to increase the infiltration capacity during the dry summer.

Hortonian overland flow, i.e. flow on the ground surface in non-saturated areas, has generally been measured by troughs in pits, oriented orthogonally to the slope direction, collecting the flow either from artificially demarked plots or from natural micro-basins. In most studies performed in vegetated, humid areas, Hortonian overland flow was absent or very small (Hursh and Hoover, 1941, in the eastern United States; Hesmer and Feldman, 1953, in West Germany; Whipkey, 1965, in the eastern United States; Ragan, 1968, in the eastern United States; Dunne and Black, 1970a, b, in the eastern United States; Weyman, 1973, in south-eastern England; Beasley, 1976, in the southern United States; Niemczynowicz, 1977, in southern Sweden; Bren and Turner, 1985, in south-eastern Australia; Ando et al., 1983, in Japan; Heede, 1984, in the south-western United States). In many studies (e.g. Corbett et al., 1975) it is stated, without reference to measurements, that no Hortonian overland flow occurs in the basin. Selby (1973) showed, from 20 runoff plot experiments in New Zealand, that the generation of Hortonian overland flow is closely related to land use. The largest producer of overland flow was pasture land, where about 5% of two years' precipitation was collected. From ungrazed grass and shrub overland flow constituted about 1% of the precipitation.

Walsh (1980) reviewed studies on runoff processes in tropical areas and concluded that Hortonian overland flow is unimportant in the ever-wet tropics. It has been reported in seasonal tropical areas, mainly in cultivated land. In arid and semi-arid areas Hortonian overland flow is generally believed to be the main supplier of streamflow, but there seem to be few studies that support this statement quantitatively.

In Sweden frontal rainstorms generally have intensities of less than 5 mm h^{-1} and heavy convective showers may deliver $30\text{--}50 \text{ mm h}^{-1}$ for a few minutes. About 50% of the rain falls with intensities lower than 2 mm h^{-1} and 80% lower than 7 mm h^{-1} (unpublished data from the Swedish Meteorological and Hydrological Institute, see Grip and Rodhe 1985). Snowmelt rates are low as compared to rainfall intensities. In Sweden rarely more than 30 mm of meltwater is produced in one day, corresponding to a maximum intensity in day time of less than a few mm h^{-1} . Lundberg (1974) reviewed infiltration measurements carried out in Sweden, Norway and Finland. From the infiltration

measurements reported by Lundberg, the only generation of overland flow on unfrozen ground would occur from heavy convective showers on certain clay soils. Later infiltrometer measurements, mainly in south-western Sweden, have been reviewed by Lindblad (1981). She reported small differences between soil types, with a log mean for till soils of 20 mm h^{-1} and for clay of 6 mm h^{-1} .

The above reports concern infiltration into unfrozen soil. Studies of infiltration into frozen soil have shown that the infiltration capacity is generally reduced and that the reduction is highly dependent upon the frozen water content of the soil. Kane and Stein (1983a) in Alaska, and Price and Hindre (1983) in Canada, observed overland flow from runoff plots only during those springs that were preceded by unusually wet autumns. Ring infiltrometer measurements by Kane and Stein (1983b) in Fairbanks silt loam showed infiltration capacities of about 15 mm h^{-1} for unfrozen soil. When the soil was frozen with low total moisture content the infiltration capacity was only moderately reduced, to about 5 mm h^{-1} . But when the total moisture content was high, the capacity declined to 0.2 mm h^{-1} after freezing. High infiltration capacities, $10\text{--}30 \text{ mm h}^{-1}$, were reported by Haupt (1967) from plot experiments with simulated rainfall on frozen ground in the western United States. In some cases the infiltration capacity increased with freezing of the soil, attributed by Haupt to favourable changes of the soil structure. Granger et al. (1984) determined infiltration in various soils (sand to heavy clay) in the Canadian prairies by soil density monitoring. At many sites with uncracked soil the total estimated infiltration, i.e., the increase in ice and water content of the soil, was considerably less than the total meltwater production. No measurements of overland flow were reported, however.

The only studies showing Hortonian overland flow as an important contributor to stream stormflow in humid areas were performed on frozen ground. Dunne and Black (1971) collected overland flow from a 50 m hillslope with frozen pasture ground in the north-eastern U.S. The water was intercepted by long troughs in a 100 m trench at the foot of the hillslope. The sandy soil is reported to have an infiltration capacity in summer larger than 80 mm h^{-1} , allowing all summer rainfall to infiltrate. During the investigated melting of a snow cover of 65 mm water equivalent, which took place during one week, the streamflow generation from the slopes was dominated by Hortonian overland flow. The reduced infiltration capacity was attributed to concrete frost on the surface and in the upper few decimetres

of the soil, caused by repeated melting and freezing earlier during the winter. A complete dominance of Hortonian overland flow in the runoff from plots during snowmelt was reported by Price et al. (1978) from a study in the Canadian Subarctic. They measured the flow from 4 plots, with areas of about 2000 m², on forested clayey glacial till that was frozen by concrete frost to a few metres depth. The water volumes collected in the surface troughs during melting corresponded closely with the total amount of snow water equivalent on the plots before melting.

2.1.2 Subsurface flow

Direct measurement of subsurface flow during rainstorms has been undertaken in natural slopes by the use of pits with troughs inserted in the upslope wall (Hursh and Hoover, 1941; Whipkey, 1965; Ragan, 1968; Dunne and Black, 1970a, b; Beasley, 1976; Bonell and Gilmour, 1978; and Mosley, 1979.) The measurements are complicated by the fact that the construction of a pit changes the flow pattern (Whipkey and Kirkby, 1978). In the case of unsaturated flow, the flow lines are diverged by the presence of the pit, resulting in an underestimation of the flow. The opposite effect takes place for saturated flow, in which case the flow lines are converged towards the pit. (Strictly speaking, the only flow that can be collected with a trough is saturated flow, because the water must be of a pressure greater than that of the atmosphere in order to leave the soil. But in the case of unsaturated flow, a saturated zone might develop above the pit and act as a transmitter of some of the unsaturated flow from above.) Errors due to horizontal convergence or divergence are reduced by using long pits. In a few cases, saturated subsurface flow from hillslope segments has been measured with troughs in natural seepage faces (Weyman, 1973; Megahan, 1983). Subsurface flow has also been measured in large inclined soil columns, draining at the downslope wall (Hewlett and Hibbert, 1963; Nutter, 1971).

In a laboratory experiment of saturated flow, Anderson and Burt (1978a) showed that if water is collected by troughs at different depths there is a great risk of drawing incorrect conclusions about the relative magnitudes of flow in different layers. At the vertical face in the pit the upper layers will be partly drained by the deeper layers, which gives an overestimation of the flow from the deeper layers.

The study by Hewlett and Hibbert (1963) in the eastern United States demonstrated that unsaturated flow was able to sustain baseflow via a saturated zone interconnected to the stream. In the same area Nutter (1971) compared the outflow from a 61 m long artificial slope segment to the stream hydrograph and concluded that unsaturated flow was the main component of stream stormflow. Early pit measurements by Hursh and Hoover (1941) in the same area indicated subsurface stormflow in the top 30 cm of the soil during natural conditions.

The results obtained in the eastern United States by Whipkey (1965), whose trough system for flow measurements on several depths has been used by many others, also indicate that subsurface stormflow may be a major component of stormflow, although the study used artificial rain and thus could not be related to the stream hydrograph. The recorded flow took place above the transition from permeable sandy loam to a less permeable loam at 0.9 m depth.

Pit measurements by Beasley (1976) in the southern United States recorded subsurface stormflow from the upper, as well as from the lower, part of a hillslope. The rapid response to rainfall led him to the conclusion that the flow took place in macropores which, due to low conductivity of the soil matrix, could be waterfilled, although the soil was not saturated. Similar results were obtained by Mosley (1979) in New Zealand. In Mosley's study of a 0.3 ha basin with 0.5 m soil on impermeable bedrock, the flow was measured in several pits and natural seeps, and related to the stream hydrograph. Flow in macropores seemed to occur both during saturated and unsaturated conditions in the soil mass. (An extreme case of subsurface stormflow is the flow in so called soil pipes of several centimetres in diameter, which form natural underground drainage systems in certain soils in England (Jones 1978).)

Weyman (1973, in south-western England) installed troughs at different depths into a natural exposed soil face at a slope base. No overland flow was recorded, and the hillslope hydrograph was dominated by saturated subsurface flow which could be related to the vertical and lateral extension of a saturated zone within the soil. The flow occurred above a level in which there was a sharp downward decrease in the saturated hydraulic conductivity of the soil. The hillslope hydrograph was not compared to the basin hydrograph, and the author did not regard it as subsurface stormflow.

The works by Dunne and Black (1970a, b; 1971) were carried out in a hillside with about 0.5 m highly permeable sand resting on a layer of varved silt underlain by silty clay till. A 100 m long trench at the base of the slope was equipped with troughs to collect overland flow, flow in the root zone, and flow in the upper layer of the permanently saturated zone. Three segments of the slope were monitored separately within the 100 m trench: one convex, one concave, and one straight slope. Natural and artificial rainstorms were investigated, and the flows were related to the stream hydrograph. Although some rapidly responding subsurface flow was recorded in the root zone its significance to stream stormflow was regarded as small. Saturation overland flow from the foot of the concave slope segment was found to dominate the stream hydrograph during storms. The authors strongly deny the existence of subsurface stormflow, but estimate the measured saturation overland flow during one storm to be about 25% direct rainfall with the remaining 75% being return flow, i.e., subsurface flow escaping from the soil before it reaches the stream. It seems a little strange, however, to emphasize a difference between "return flow" and "subsurface flow", the latter term being restricted to water in the soil discharging directly into the stream. As far as I understand, these studies by Dunne and Black show that stream stormflow was dominated by contributions from soil water or groundwater discharging outside the stream. Ragan (1968), whose study was made in homogeneous sand in the same region, recorded no subsurface flow in his pit, except for a flow in the litter during heavy storms, estimated to be about 10% of the hillslope contribution.

In an Australian tropical rain forest, pit measurements in a 0.3 km² basin with steep slopes were made by Bonell and Gilmour (1978) and related to the stream hydrograph. The soil is highly permeable in the top 0.2 m, below which depth the hydraulic conductivity decreases over a short distance. In this area, with an annual rainfall exceeding 4000 mm, a perched groundwater in the top 0.2 m developed rapidly following rainfall. Thus the whole area became saturated, causing saturation overland flow from the whole basin. The measured subsurface flow was of minor importance for the stream stormflow hydrograph, which was considered to be dominated by saturation overland flow.

The above-mentioned attempts to measure subsurface flow were all carried out in areas where the permanent water table did not reach above the top metres of the soil, except in the close vicinity of the stream. In those cases where rainfall was

reported to cause a substantial increase of subsurface flow (though not always considered as subsurface stormflow), it occurred in the top 0.5 m of the soil, either as saturated flow in an occasionally saturated permeable layer overlying a layer of low permeability (Hursh and Hoover, 1941; Whipkey, 1965; Weyman, 1973; Dunne and Black, 1970a, b) or as flow in macropores in unsaturated as well as saturated soil (Beasley, 1976; Mosley, 1979). The only case where no subsurface flow (except in the litter) was observed at pit measurements was in the study by Ragan (1968), whose measurements were made in homogeneous sand. Thus, a large decrease in the hydraulic conductivity (either textural or structural) with depth seems to be a prerequisite for the occurrence of subsurface stormflow along slopes with low permanent water tables.

2.1.3 Soil water and groundwater potentials

Since the hydraulic conductivity in soils often increases by orders of magnitude in a small tension interval when going from unsaturated to saturated conditions, the existence of saturated zones has a decisive role in the water flow. Measurements of subsurface water potentials and moisture content in the studies on the runoff process have given information about the direction and magnitude of subsurface flow. Important results concerning the role of topography have been achieved in some studies.

Lundin (1982) made a detailed investigation of soil and water conditions in a small basin with glacial till soil in central Sweden. He observed a characteristic depth variation of the saturated hydraulic conductivity, which at many sites decreased by two orders of magnitude from 0.1 m to 0.5 m below the ground surface. Corresponding depth variations were also observed for the total porosity (decreasing with depth in the top layers) and for the water-holding capacity of the soil (increasing with depth). From long-term observations of groundwater levels and soil moisture content and from tracer experiments, he showed that lateral groundwater flow in the topmost layers of the soil was the main component of stream stormflow. The magnitude of the groundwater flow was very sensitive to variations in the groundwater level. The results of Lundin, which are frequently referred to throughout this thesis, are discussed in more detail in Chapter 10.

Betson et al. (1968) installed maximum-level recording piezometers in the A-horizon (0.1 - 0.3 m deep) in a basin with deep

permanent water table. The existence of an occasionally saturated layer during storms was demonstrated, predominantly in areas with a thin A-horizon. The saturated flow in this layer in the upslope areas, dominated by a shallow A-horizon, was considered to contribute to stream stormflow only during heavy storms, because of subsequent deep layer infiltration downslope. From piezometric measurements in a small, gently sloping depression in Luxembourg, Bonell et al. (1984) concluded that storm runoff appeared when the perched groundwater reached the top few cm of the soil, having high hydraulic conductivity due to biological activity. The lateral flow in macropores in this layer was considered more effective in generating stormflow than was saturation overland flow, which also occurred.

From long-term soil moisture measurements along hillslopes with a deep water table, Halvey et al. (1972) showed that the soil moisture content increased down the slopes, thus giving evidence to the notion of near-stream areas as source areas for streamflow (by saturated or unsaturated flow initiated by local infiltration). High soil moisture content close to the stream, interpreted as high potential for generation of surface runoff, was also shown by Henninger et al. (1976), who investigated the spatial variability of surface soil moisture within an agricultural watershed.

Changes in soil water potential along a slope with a gauged natural outlet were studied by Weyman (1973). The measured outflow could be related to the vertical and upslope extent of a wedge-formed saturated zone at the slope base. The importance of the saturated zone within the soil at the slope base was also pointed out by Harr (1977), who, in a very instructive paper, analysed the flow in a steep 110 m long, highly permeable forested hillslope (gradient 75%) in a 0.1 km² watershed (Oregon, the north-western U.S.). Soil depth was 3-8 m, with a permeability decrease by a factor 10 at 1.2 m depth. Flow directions were determined from tensiometer recordings, and flow magnitudes estimated using empirical relationships between measured saturated conductivity and unsaturated conductivity as a function of the water tension. During storms, unsaturated lateral flow dominated in the upslope parts of the hillslopes, although saturated zones of limited areal extension developed at a depth of 1.1-1.5 m. These saturated zones, however, seemed to be isolated by unsaturated regions, which regulated the downslope water flow. In the lowest 12-15 m of the slope, a saturated zone within the soil transmitted the unsaturated flow from above and upslope. The storm discharge peak (quickflow)

constituted on the average 38% of the rainfall over the basin. Out of this 38%, channel interception constituted 3%, and subsurface stormflow 97% (no overland flow was observed). The fact that as much as 38% of the rainfall over the basin ran off as quickflow, indicated that a considerably larger portion of the slope than the 12-15 m wide zone along the stream, in which there was a watertable, contributed to the quickflow. The results stress the importance of unsaturated lateral flow in these slopes.

By detailed monitoring of soil water potentials in the lower part of a hillslope of 1-2 m deep, highly permeable soil, resting on impermeable rock, Anderson and Burt (1978b) were able to relate water potentials and subsurface flow to the micro-topography. Spurs and hollows in the hillslope are oriented orthogonally to the stream. The flow was always converging to the monitored hollow, where a saturated zone within the soil, sometimes extending to the surface, rapidly developed during rainstorms. The surrounding spurs remained unsaturated except during very intense storms. Because of small areal differences in tension potentials, the total potential gradient was determined mainly by the topography. Flow direction could thus be determined from a topographical map. The depth of the saturated zone could be related to stream stormflow, which showed a shortlived, rapidly developed peak attributed to rainfall on the stream (channel interception), and a delayed peak caused by subsurface stormflow. The contribution to streamflow per unit ground area, estimated from discharge measurements along the stream, was about twice as large from the hollows as from spur areas. In a later study, Anderson and Kneale (1980) used the same method to examine the flow conditions in a more gentle hillslope (sloping 6% as compared to 25% in the earlier study) of less permeable, pastured clay loam. In this case, the total potential gradient could not be simply related to the topography. A saturated zone, fed by flow from the instrumented hollow, developed in the downstream spur. The smaller degree of control by topography was attributed to the smaller height differences, which in this case were comparable to areal differences in water tension.

The influence of three-dimensional topography on the water flow was also demonstrated by Dunne and Black (1970a, b), who found the main contribution to storm hydrographs from the convergent slope segment.

Field mapping of saturated areas, on basin-wide scales, has been reported by Dunne et al. (1975) in various areas in North America, Beven (1978) in England, Allison and West (1979) in southern Australia, Taylor (1982) in Canada, and Myrabö (1986) in southern Norway. The extent of saturated areas was determined at different discharges in the streams, showing increasing saturated areas with increasing streamflow. Myrabö found that the rainfall on the saturated areas corresponded to about half of the quickflow observed in the stream and concluded that there was a considerable subsurface component of stream stormflow. Dunne and Black (1970a, b) and Taylor (1982), on the other hand, found that the amount of rainfall on saturated areas roughly equalled the volume discharged as quickflow in the stream, indicating little contribution from subsurface flow (or return flow). Allison, working in a pastured basin with soils of low hydraulic conductivity, obtained results similar to those of Dunne and Black (1970a, b) and Taylor (1982) at low rainfall intensities. At moderate and high intensities, however, the discharged volume exceeded the amount of rainfall on saturated areas. The difference was explained as a contribution of Hortonian overland flow from the remainder of the basin, a flow that was also observed qualitatively. Beven, on the other hand, found that in some basins the rainfall on saturated areas exceeded the discharged volume. This discrepancy was explained by the fact that certain areas were saturated without being connected to the channel network. In other nearby basins, lengths of flow paths and velocities of overland flow suggested a considerable contribution from return flow to the hydrograph peak and early recession.

2.1.4 Other experiments

The origin of subsurface stormflow was investigated in a basin-wide irrigation experiment by Corbett et al. (1975). The average sideslope steepness of the 0.8 km² forested basin is 30%, and soils are mostly silty loam, 0.5 m deep at the upper slope, and 1-2 m deep in the channel area. Storm hydrographs in the draining stream resulting from sprinkler irrigation of different parts of the basin were analysed. Combinations of three portions of the basin, the channel area, the lower slopes, and the upper slopes together with the ridge area, were irrigated at a rate of 6 mm h⁻¹ for 8 h at times when the basin was near field capacity. Comparisons of the stormflow hydrographs showed that stormflow during the first 2-2.5 h was initiated exclusively from the channel area. After two hours, subsurface flow from

the lower slopes was added and after two more hours, i.e., four hours after the beginning of the irrigation, subsurface flow from the upper slope started to contribute. The combined flow from the channel area and the lower slopes peaked just after the cessation of the artificial rainfall, while the flow from the upper slope and ridge areas continued to increase during more than three hours, resulting in total basin peak two hours after the cessation of rainfall. The experiment clearly showed the importance of subsurface stormflow also from the upper slope area in this basin.

2.2 Simulations with mathematical models based on Darcy's law

The understanding of the processes involved in streamflow generation is developing in a dialectical way between field measurements and mathematical modelling. On the one hand, the mathematical model (which may be the simplest of relationships or a complicated system of equations) is formulated from knowledge about the processes in nature. On the other hand, results from the model will aid in interpreting the results obtained by field measurements and make it possible to generalize the results. Mathematical modelling of hillslope runoff has mainly been used to examine how certain measurable parameters, such as saturated conductivity and topography, influence the production of runoff.

The basic equations for hillslope models are Darcy's law and the equation of continuity. Since the complete solution of the unsaturated-saturated flow caused by an infiltration event requires considerable computer time, various simplifications are made. In some models the flow is assumed to be vertical in the unsaturated zone, or the groundwater is assumed to be recharged with a delay depending only on the depth to the water table. Some authors simplify the saturated flow by use of the Dupuit assumption, i.e., by assuming horizontal flow or flow in parallel with impermeable bottom of the flow region only. The authors differ in their numerical methods of solution and, of course, in their various applications of the models.

The first complete saturated-unsaturated hillslope flow model was presented by Freeze (1971), formulated in three dimensions but run in two dimensions. From simulations reported in later papers (Freeze 1972a, b), he concluded that subsurface stormflow could be of importance only in steep, convex hillslopes with soils of extremely high saturated conductivity, and that

overland flow from surface saturated areas close to the stream dominated the flow in all other cases. There was no significant contribution to stormflow from the upper slopes. The model was applied to a field situation during snowmelt by Stephenson and Freeze (1974). The soils in the 250 m hillslope segment, which has a gradient of about 25%, rest on a layer of fractured basalt. This stratigraphy gives an increase in saturated conductivity with depth, down to the impermeable bedrock. Parameters were calibrated against measured groundwater levels and the basin outflow hydrograph. No overland flow was observed or modelled. They modelled the flow with "neither particularly good nor particularly bad" agreement between measured and model conductivities and concluded that the peak runoff was caused exclusively by infiltration in the lower part of the slope. The authors pointed out that even in this comparatively well instrumented hillslope, there was a large gap between the data needed for describing the flow by the model and the available field data.

A similar unsaturated-saturated hillslope model, solved by the finite difference method, was used by Beven (1977). The simulations were performed for a 100 m hillslope of 20% gradient, with one metre of homogeneous soil on impermeable bedrock. The initial conditions were extremely dry, with drainage equilibrium assumed for the whole slope, i.e., saturation at the bottom of the base, and negative pressure potential in the slope equal to the height above the bottom. Consequently, the simulated subsurface flow was very small and delayed, with peakflow draining only a few per mil of the applied rainfall amount. The only considerable lateral flow took place in the saturated zone, which developed from the base about 10 m upslope, never reaching the surface. This zone was fed mainly by vertical percolation, not by lateral unsaturated flow from upslope, a result which was considered to support the variable source area concept. Simulations with different topography nevertheless stressed the importance of slope convergence in the development of a saturated zone at the base.

In order to explain the large groundwater contributions to stormflow assessed by their isotopic studies (see Section 2.4), Sklash and Farvolden (1979) simulated the flow in short (9 m) hypothetical slopes with soils of constant saturated conductivity, using an unsaturated-saturated hillslope flow model allowing seepage surface. Groundwater flow dominated the stormflow in the simulated events. In agreement with a few field measurements, the groundwater levels close to the stream rose

rapidly during storms, increasing the gradient and thus the groundwater discharge into the stream. According to the simulations, which started with a horizontal groundwater surface, the rapid response close to the stream allowed a groundwater ridge to develop. Comparisons of slopes of various lengths indicated that the upper parts of the slopes did not contribute to stormflow in the simulated storms.

Johansson (1985) applied an unsaturated-saturated flow model to hillslopes intended to represent conditions in Swedish till soils. An impermeable bedrock was covered by 1 m of vertically layered soil, with hydraulic conductivity and porosity increasing towards the ground surface according to the observations by Lundin (1982). When the initial moisture condition in the slope corresponded to a steady-state flow of 0.07 mm h^{-1} , simulated infiltration of 5 mm h^{-1} for 6 h generated a groundwater outflow hydrograph similar to observed streamflow hydrographs in small till basins. According to the simulations, the response of the groundwater outflow was mainly due to infiltration close to the stream. There the groundwater table rose into shallow layers of the soil having high conductivity, which, combined with an increased gradient of the water table, drastically increased the outflow. With the applied rainfall intensities, surface saturation did not occur in the straight hillslope, but in simulations for a similar but concave slope a saturated area soon developed.

Beven (1981, 1982a, b) used a kinematic wave approximation to simplify calculations of subsurface stormflow. The model was applied with conductivity values reported in the literature from field studies where subsurface stormflow was observed. He concluded that subsurface stormflow could be generated only in soils of very high hydraulic conductivity, where the groundwater table or an impeding layer is close to the ground surface. But these conditions are, as Beven noted, met in many areas.

Another simplified hillslope model was developed by Bren (1980) and applied with good results to describe streamflow in a small first-order stream in Australia. No saturated areas were observed in the basin, and overland flow was insignificant. In the model the groundwater flow was described with the Dupuit approximation and fed by infiltrated rainfall with a time lag proportional to the depth of the water table, which in turn was assumed to be proportional to the distance from the stream.

Eriksson (1977) analysed steady-state flow situations in a straight hillslope by the Dupuit assumption and discussed the extension of saturated areas as a function of the percolation rate, hydraulic conductivity and slope angle. According to these calculations, water-logged areas should frequently develop in till soils, since the flow capacity of the lower parts of many slopes can be expected to be exceeded by the water flow from upslope.

From similar considerations O'Loughlin (1981) derived criteria for the existence of a surface saturated zone at the slope base during steady-state flow, and related the extension of the zone to soil conductivity and topography. The observed volumes of quickflow from a few Australian basins agreed approximately with the amounts of rainfall on surface-saturated areas, the extensions of which were estimated from the model. The contribution of subsurface flow was thus considered small.

The large areal variability of saturated conductivity introduces a difficult problem in hillslope flow modelling. The problem was tackled by Freeze (1980) in a stochastic-deterministic model for calculation of overland flow. Soil properties and rainfall characteristics were treated statistically, while the flow in the soil was calculated with physically-based equations. Saturated conductivity was assumed to be log-normally distributed. During the simulated rainfall events, both Hortonian and saturation overland flow occurred on the slope with the dominance of either mechanism, depending on soil/rainfall parameters and on the slope. The simulated overland outflow hydrographs were reported to be similar to some overland flow generated peaks reported in literature. The author pointed out that large errors might arise when a heterogeneous hillslope is treated as homogeneous, with some type of mean value of saturated conductivity.

In most field studies and mathematical models of runoff processes, the slopes have consisted of - or have been supposed to consist of - a thin soil cover (a few metres) on impermeable bedrock. Consequently, the saturated flow direction has been mainly lateral. The vertical components of groundwater flow, on the other hand, were emphasized by Tóth (1963), Freeze and Witherspoon (1967), and Gustafsson (1968). They modelled groundwater flow in deep vertical cross-sections of landscapes with undulating groundwater surface. These analyses focused on the existence of different flow systems, with recharge and discharge areas for groundwater generated by the topography. Their results provide a very important basis for the understanding of ground-

water flow within a basin, but since they treated only stationary conditions, little direct information was given on the dynamics of stormflow generation. It should also be noted that their flow patterns were calculated for very thick aquifers having constant hydraulic conductivity (Tóth, Gustafsson), or for aquifers consisting of thick layers of constant permeability (Freeze and Witherspoon). The effect of a large increase in conductivity close to the ground surface, as reported from many field studies, is thus not seen. But these analyses show that in areas with shallow groundwater there is a contribution to streamwater from deep groundwater, even though the flow rates may be small. The implicit or explicit assumption in most studies, that the underlying bedrock could be treated as impermeable, might therefore lead to erroneous conclusions, for instance when chemical considerations are made. The role of deep groundwater flow in streamflow generation in till soils on crystalline fissured rock is little known.

2.3 Interpretation of streamwater chemistry or isotope content

The field studies referred to in the preceding sections mostly illuminate one or a few of the several processes contributing to streamflow generation. Analyses of the chemical or isotopic composition of streamwater during runoff events, on the other hand, give information on the integrated result of the various processes. The method of hydrograph separation by isotopes, which is the basis for the present study, is described in Chapter 3; here some comments are made on applications made by other authors.

Actually, all attempts to explain measured data on streamwater chemistry rely on some particular view of the runoff process, implicitly or explicitly. Qualitative conclusions about the runoff process from interpretation of streamwater chemistry have been drawn in several studies. These analyses utilize different chemical labelling of the water, which takes place at various localities within a basin. The precipitated water, poor in solutes, becomes gradually enriched as it passes through the basin. The final characteristics of the water entering the stream depends on the flow paths and transit times. Temporal changes in streamwater chemistry during runoff events, or spatial changes along a stream, are interpreted as changes of the proportions of water of different origin. An inverse relationship is often found between the concentration of certain chemical species or total salt concentration and discharge. The

assumption is then that highly concentrated groundwater is diluted by less concentrated rain- or meltwater.

A method for quantitative estimates of the various flow components in streams was derived by Ivanov (1948), from continuity conditions for water and solutes. By a modification of the method, Zekster (1963) estimated the contribution of deep groundwater to the river Nema in the eastern U.S.S.R. to be about 20%, using chloride as a tracer. In two Japanese studies, radiostrontium and radiocesium, emitted into the atmosphere by hydrogen bomb tests, were used to trace the origin of streamwater. The few data shown by Yamagata et al. (1963) suggest a large fraction of groundwater according to the $^{89}\text{Sr}/^{90}\text{Sr}$ ratio in different waters. From comparisons between ^{90}Sr and ^{137}Cs (the latter isotope being rapidly adsorbed by the soil) Miyake and Tsubota (1963) estimated the fraction of rainwater precipitated directly in the stream (channel interception) to be <2% of total streamflow. Pinder and Jones (1969) performed hydrograph separation by the use of a composite of several major ions in three small glaciated till basins in south-eastern Canada. According to their estimates, 32-42% of the discharged peak volumes originated from groundwater.

The electrical conductivity, being a measure of the total salt concentration, has frequently been used in chemical hydrograph separation. In a central U.S. basin, Visocky (1970) estimated the groundwater contribution to be about 25% from measurements of conductivity. Tagutschi (1982) shows groundwater fractions from about 30 to 80% in two Japanese basins. The dominant role of throughflow in a small British basin was shown by Anderson and Burt (1982). They compared the results of hydrograph separation by electrical conductivity to the results of detailed hydrologic measurements (Anderson and Burt 1978b, referred to in Section 2.1.3). Both methods showed the dominance of throughflow, but the differences they obtained led the authors to issue a warning against interpretation of chemical data collected at a few sampling points in terms of details of the hydrological processes within a basin.

Large groundwater fractions in snowmelt runoff in a Japanese basin were obtained from conductivity measurements by Kobayashi (1986). Based on the finding that the discharging groundwater was warmer than the meltwater, Kobayashi also used water temperature as a tracer, thereby obtaining similar results. The temperature could be used as a tracer as long as the heat

exchange between streamwater and the atmosphere was reduced by an ice-cover on the stream.

When only one tracer is used, the hydrograph can be separated into only two components. Three-component separation was performed by Balek et al. (1978), who separated the hydrograph into surface flow and interflow using electrical conductivity and calculating baseflow from measured groundwater levels. Newbury et al. (1969) interpreted the groundwater contribution to a few discharge peaks in a basin in southern Canada to be about 50% from observations of conductivity. From SO_4 measurements they further estimated the contribution from deep groundwater, rich in SO_4 , to be about 30%.

One problem in chemical hydrograph separation is to obtain a realistic value for the tracer concentration of the rainwater flow component. Rainwater reaching the stream as overland flow may, as pointed out by Nakamura (1971), Fritz et al. (1976), and Pilgrim et al. (1979), be enriched in salts by contact with the soil particles. In the studies referred to in this section, the composition of the rainwater component has been assumed to be equal to that of rainfall, to the composition measured in a few overland flow collectors, or to that of streamflow at the time of maximum flow of the largest event, when no groundwater contribution was expected. (The latter assumption seems to seriously prejudice the results.)

In some studies a slight increase in the salt concentration of streamwater has been observed at the beginning of the runoff events before the decrease takes place. The existence of such a "salt peak" shows that the chemograph at the early stage of an event cannot be explained by dilution only (e.g. Miller and Drever, 1977), and the simple two-component separation is not possible. The salt peak may be caused by flushing of salts by the first groundwater discharging outside the stream, where salts have accumulated due to evaporation before the event (Calles, 1982). In areas where direct evaporation from soil is important it could, however, also be due to flushing during overland flow on recharge areas.

The above investigations are event studies, separating the hydrograph during one or more runoff events. The groundwater contribution to streamflow has also been investigated by measurements of streamwater chemistry along streams at certain moments. Many studies have shown that the electrical conductivity increases downstream. This has been interpreted either

as the result of an increasing fraction of groundwater with long transit time (Pinder and Jones, 1969; Calles, 1985), or as groundwater contribution from different geological formations (Appelo et al. 1983). Monitoring of conductivity at several points along a Swedish stream enabled Calles to estimate variations in the deep groundwater contribution along different reaches of the stream during runoff events. According to her calculations, there was a rapid and large increase also in the deep groundwater flow to the stream during stormflow.

The use of water chemistry as a tracer in runoff studies relies on the non-conservative property of water chemistry during water flow through a basin. The streamwater is considered to be a mixture of water that has undergone various chemical changes, depending on the flow paths to the stream. The use of environmental isotopes of the atoms of the water molecules, on the other hand, is based on the conservative behaviour of the isotope content of the water during the process of runoff. In this case, hydrograph separation and estimates of transit times are possible due to the time variation of the isotopic composition of the precipitation and the damping of this variation during water flow through various reservoirs within a basin.

Hydrograph separation by isotopes has been performed since the early 1970s, predominantly in mountainous areas in central Europe and in Canada. Runoff events generated both by snowmelt and by rainfall have been investigated.

Crouzet et al. (1970) used tritium to estimate the groundwater component during rainfall in three streams in south-eastern France, and showed examples of discharge peaks dominated by either groundwater or direct precipitation. Dinger et al. (1970) analysed the runoff from the Modro Dul basin in northern Czechoslovakia using tritium and ^{18}O . They concluded that about 70% of the snowmelt runoff originated from groundwater. In similar studies in the Ditchma basin in Switzerland, Martinec et al. (1974) found that groundwater contributed about 60% of the mean runoff. Snowmelt and rainfall generated runoff was also investigated using tritium and ^{18}O by Herrmann et al. (1978) and by Herrman and Stichler (1980) in a very detailed study in the Lainbach valley in the Bavarian Alps in southern West Germany. In this basin, too, groundwater dominated streamflow during peaks. According to their estimates, groundwater contributed about 70% of the mean runoff from the basin.

In a study of several Canadian basins using ^{18}O and water chemistry, Fritz et al. (1976) found that groundwater (or pre-event water) constituted 40 to 90% of the discharged volumes during rainfall-generated runoff events. Further studies of rainfall events in Canadian basins reported by Sklash and Farvolden (1979) and by Bottomley et al. (1984) show similar results. Large groundwater components during rainfall were also reported from isotope studies by Mook et al. (1974), analysing a runoff event in a Dutch basin, and by Blavoux (1978) and Merot et al. (1981) analysing runoff events in French basins. Other isotope studies of snowmelt events, also showing the dominance of groundwater or pre-event water, are those by Bottomley et al. (1986) in Canada and by Hooper and Shoemaker (1986) in Hubbard Brook in New Hampshire, U.S. Hooper and Shoemaker discuss important methodological problems involved with hydrograph separation of snowmelt events, such as the time variation of the meltwater and pre-event water components.

In contrast with the results of other isotope studies, Jacks et al. (1986) found from ^{18}O observations that streamflow during a spring snowmelt was almost entirely new meltwater. The 34 km² basin is situated in Härjedalen, central Sweden at 600 to 900 m above sea level. About half of the basin is sloping boglands and the remainder is covered by forest. Hydrograph separation was not performed, but after one week of hydrograph rise, when stream discharge had reached half of its maximum value, the isotopic composition of meltwater (from lysimeters) was almost the same as that of streamwater. The similarity remained until the lysimeters were snowfree. A suggested explanation for the dominance of meltwater, shown also by pH-measurements, was the contribution of saturation overland flow from the bogs, which got saturated to the surface after a few days of melting.

Pearce et al. (1986) and Sklash et al. (1986) performed a detailed isotopic study of runoff in the same small highly responsive basin in New Zealand as the one in which Mosley (1979) made tracer experiments and flow measurements in pits (cf. Section 2.1.2). Several rainfall-generated runoff events were investigated in the basin and its neighbouring basins, all of which showed large pre-event components of streamflow (75-97% of the total flow volume). The earlier conclusion by Mosley, that the rapid subsurface contribution to streamflow was due to rainwater flow in macropores, was refuted by the isotope study. According to the isotope data by Sklash et al., the subsurface flow collected in pits in the hillslopes was mainly pre-event water. They attributed the rapid response of

the basin to the effect of the capillary fringe (Gillham 1984), by which large head gradients towards the stream can rapidly build up in the riparian areas.

Hydrograph separation into more than two components was performed by Behrens et al. (1971, 1979) using tritium, deuterium, and electrical conductivity in three partly glaciated Austrian basins. Griend, Van de, and Arwert (1983) concluded from ^{18}O measurements that stored meltwater dominated a rainfall-generated discharge peak in a partly glaciated basin in northern Italy.

The simple models underlying the hydrograph separations presented in this section were questioned by Kennedy et al. (1986). They analysed chemical and isotopic data from a group of very large rainfall events (250 mm of rainfall in 6 days) in a 620 km² basin in north-western California, U.S. As in most other studies, the chemistry and isotope content of streamwater showed little influence of new rainwater. Depending on the tracer used, pre-event fractions of 50-80% of total streamflow were obtained from two-component separations. According to Kennedy et al., such large groundwater contributions are impossible in this basin, considering the low hydraulic conductivities of the soil and bedrock and the deep water table. (Very little quantitative support for this opinion is given in the paper, however.) The apparent high contribution of pre-event water is attributed by the authors to water exchange between overland flowing water and soil water. When rainwater is flowing as (Hortonian) overland flow, it infiltrates into the soil at certain localities. The infiltration is followed by exfiltration of pre-event soil water a few millimetres to a few metres downslope. Thus, according to their explanation, the overland flow reaching the streams may have the isotopic composition of the pre-event soil water, which may differ from that of the groundwater. The overland flowing water may further have been subject to chemical reactions in the soil.

In several of the central European studies mentioned above, mean transit times of subsurface waters were estimated from the observed temporal variations of the isotope content of precipitation and streamwater, (Dinger et al., 1970; Martinec et al., 1974; Maloszewski et al., 1983; and Behrens et al., 1979). Mean transit times around 2-4 years were computed, with differences between basins, and also for the same basin depending on the assumed flow models. Pearce et al. (1986) estimated the mean

transit time for the "old water component" of streamflow in their New Zealand basins to be 4 months.

In most of the runoff events for which hydrograph separation by means of isotopes has been reported, groundwater dominated the total discharged volume. Less than 40% groundwater at rainfall events was reported in only one of the three events investigated by Crouzet et al. (1970), and in one of the several events investigated by Sklash and Farvolden (1979). The former event was generated by a rainfall of moderate intensity, while the latter was generated by a high intensity rainfall (30 mm h^{-1}). Groundwater also played a minor role in generating the runoff from the partly glaciated areas studied by Behrens et al. (1971) and in the snowmelt runoff from a Swedish basin with a large bogland area reported by Jacks et al. (1986). Of particular interest for Swedish conditions is the study by Sklash and Farvolden (1979). Some of their investigated Canadian basins have a climate, geology, and groundwater levels which make them comparable to many Swedish basins. The groundwater component was reported to constitute up to 80% of the total discharged volume during the investigated events.

The hydrograph separations using isotopes have generally yielded larger groundwater fractions than those using water chemistry. In most studies, only one of the methods has been used for quantitative estimates, either isotopes or water chemistry. The difference may thus reflect different flow conditions in the basins investigated. It seems, however, more probable that the difference is fictitious, caused by methodological problems in the interpretation of streamwater chemistry, and in some cases by different definitions of the flow components. Fritz et al. (1976) compared the isotopic and chemical changes of streamwater during some runoff events. Their data indicated that the areal variability of the chemical composition of groundwater within a basin may partly explain the observed time variations in streamwater chemistry (cf. also Walling and Webb, 1980). Increased relative contribution of shallow groundwater, poor in solutes, may further dilute the streamwater (e.g., Calles, 1985), and erroneously give the impression of a large rainwater contribution. Since the tritium, deuterium, and ^{18}O content of subsurface water is changed only by mixing, while the chemical composition is gradually changed due to interaction with the soil and bedrock, the areal and depth variability of the isotope content of groundwater is generally small as compared to that of

the chemical composition. For this reason, hydrograph separation by isotopes is probably more reliable than separation by water chemistry.

2.4 Conclusions from the review of literature

The Horton model for runoff is of little relevance in humid temperate areas, since the infiltration capacity of the soil generally exceeds the intensity of rainfall or snowmelt. The only cases where Hortonian overland flow was reported to be the dominant process of runoff were the studies on frozen ground performed by Dunne and Black (1971) and Price et al. (1978).

Rainfall directly on the stream and on the surface-saturated areas connected to the stream is, of course, one of the components of stormflow. By some authors it is believed to be the sole component, giving a very restricted meaning of "contributing area" as surface-saturated area. The dynamic nature of this area is stressed in many papers. In some studies, the volume of quickflow was found to be equal to the rainfall amount on saturated areas. In such cases, the contribution from subsurface flow to stormflow should be small.

The existence of subsurface stormflow is demonstrated by some of the reported flow measurements performed in pits or natural seepage faces. It should be noted, however, that the methodological problems in such measurements are considerable, since the measurements tend to underestimate the unsaturated and to overestimate the saturated flow through the investigated cross-section of the soil.

Evidence of subsurface stormflow has also been obtained from detailed measurements of soil water potentials using measured conductivity values to calculate the flow. Since most studies have been undertaken in slopes with a deep permanent groundwater table, much interest has been focused on the development of temporarily saturated zones within the soil. In some studies, the depth and areal extension of such zones have been shown to control subsurface flow.

When subsurface stormflow has been directly observed, it has occurred in a temporarily saturated or near saturated layer with underlying less permeable soil or bedrock. In most cases, the flow has been considered to take place in macropores.

The important role played by the topography has been demonstrated by field studies as well as by mathematical modelling. Surface-saturated areas, or areas with high soil moisture content and shallow groundwater, predominantly develop at the base of concave slopes and in convergent areas. Due to water flow from upslope, such areas remain wet for a long time after a runoff event, and the flow in the areas will consequently respond rapidly to the next rainfall.

According to the isotope studies, stream stormflow is often dominated by subsurface flow. The isotope studies, however, give only the net result of the various flow processes within the investigated basins. Little information is obtained about the processes involved. There seems to be a large gap between the detailed investigations of single hillslopes and the basin input-output studies in which isotopes are used as tracers. There are no studies reported where the two types of investigation have been performed in the same basin.

Many papers discuss what portions of a slope, or parts of a basin, contribute to stormflow. Except for the very restricted definition of the contributing area as a surface-saturated area connected to the stream, the meaning of the word "contribute" is not well defined. Some authors seem to discuss the area where a considerable lateral flow occurs, others the area from where the off-running water particles originate, and others the area in which rainfall directly or indirectly affects streamflow during a runoff event. These questions, which are very important for the understanding of the chemical composition of streamwater, are discussed in Chapter 10.

3 METHODS

In the first part of this chapter, the method by which streamflow is separated into flows of groundwater and rain- or melt-water is presented. The second part gives an introduction to the basic processes which make naturally occurring oxygen-18 a useful tracer in hydrology.

3.1 Origin of streamwater traced by water chemistry or isotopes

Earlier applications of hydrograph separation by water chemistry or isotopes were presented in the preceding chapter, Section 2.4. The method relies on the mass balance of water and tracer (chemical component or isotope) during the formation of streamwater by mixing of waters of different origins. The number of flow components which can be determined exceeds the number of independent tracers by one. In the present study, streamwater during runoff events is separated into two components, using environmental oxygen-18 as a tracer. Total dissolved solids, i.e., the electrical conductivity of the water, is used as a complement in some methodological discussions.

The tracer concentration of streamwater is interpreted in terms of contribution by waters of different origin according to the model of a basin sketched in Fig.3.1.

Using the notation in Fig.3.1 the fraction of groundwater, X , in streamflow is obtained from

$$Q_s = Q_g + Q_p \quad (3.1)$$

and

$$Q_s c_s = Q_g c_g + Q_p c_p \quad (3.2)$$

which, since $Q_g = X Q_s$, give

$$X = \frac{c_s - c_p}{c_g - c_p} \quad (3.3)$$

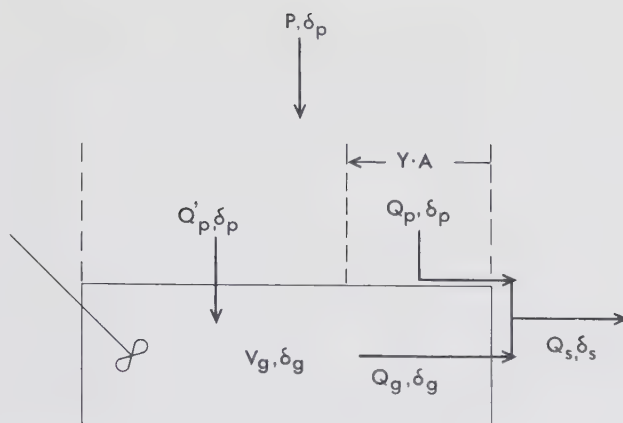


Fig.3.1 Model for interpretation of tracer concentration in streamwater.

Q_s , Q_g , and Q_p = flows of streamwater, groundwater, and precipitated water (rain- or meltwater) respectively

c_s , c_g , and c_p = the corresponding tracer concentrations

P = rain or melting intensity

Q'_p = groundwater recharge by new rain- or meltwater

A = basin area

Y = basin fraction of discharge area

V_g = volume of groundwater reservoir

Eq.(3.3) is the widely used formula for two-component hydrograph separation by water chemistry or isotopes (e.g., Pinder and Jones, 1969). The precision of the method is discussed in Chapter 9. Suffice it here to make some comments on the basic assumptions of the model and on the problem of assessing a value of c_g , the tracer concentration of the groundwater component.

The equation presumes that the stream is supplied by water from two reservoirs only, groundwater and new rain- or meltwater, each having a uniform concentration of the tracer. The separation is possible if the difference in tracer concentration between the reservoirs is large in comparison with the variability within the reservoirs.

The assumption about two reservoirs is reasonable if the participating groundwater can be treated as one reservoir with regard to tracer concentration, and if there are no lakes or other surface-water reservoirs in the basin when the runoff event starts. The contribution to streamflow from lateral unsaturated flow (soil water flow) is included in the groundwater flow.

(This may be justified from a formal point of view, since water cannot leave the soil unless its pressure is greater than that of the atmosphere. However, with the definition of groundwater given in the introduction to Chapter 2, water in saturated zones may not always be considered as groundwater.) In soils with shallow groundwater table, as those in the study basins, the distinction between soil water and surface-near groundwater is not considered important as far as the contribution to streamflow is concerned. Contribution from pre-event soil water to streamflow is therefore considered to introduce an error in the hydrograph separations only if the tracer concentration of the soil water differs from c_g , the assumed tracer concentration of the groundwater component. (See further the discussion in Section 8.1.4.)

Eq.(3.3) further presumes that the tracer is conservative after leaving the two reservoirs, i.e., that the tracer concentrations are changed only by mixing of waters of different tracer concentrations. This question is discussed in detail for ^{18}O in Chapter 6.

For some chemical compounds used as tracers, for which reaction between soil and water takes place very rapidly, the tracer concentration of the groundwater component (c_g) may remain constant with time. In such cases, c_g can be obtained from streamwater sampling before or after runoff events, when the stream is fed by groundwater only. The concentration of most chemical components of groundwater, however, gradually changes along the flow-paths due to slow chemical reactions with the soil. For this reason, the tracer concentration of the groundwater component of streamflow often decreases with increasing flow rate due to increasing relative contribution from shallow, young groundwater. After the event, when the streamwater again originates from deep groundwater, c_g may have returned to the pre-event value. Isotopically, on the other hand, rainwater¹ is transformed into groundwater by mixing with pre-event water, implying that c_g during an event may be affected by the isotopic composition of the rainwater and that there may be a net change in c_g during the event. Thus the applications of Eq.(3.3) with water chemistry and with isotopes are basically different. The following discussion refers to the use of isotopes (isotopes of the atoms of the water molecules).

¹ In these discussions the term rainwater includes meltwater.

The tracer concentrations of streamwater (c_s) and rainwater (c_p) can be measured directly from sampling. If, further, the temporal variation of the tracer concentration of the groundwater component (c_g) within a runoff event were known, then X in Eq.(3.3) would represent the true groundwater fraction of streamflow and $1 - X$ the fraction of streamflow originating from overland-flowing rainwater. Any change of c_g during the event, due to mixing with pre-event soil water or new rainwater, would be accounted for. The calculated flow of groundwater in the stream, $Q X$, would include some pre-event soil water or new rainwater that had been transformed into groundwater and discharged in the stream. However, as will be discussed in Section 8.1.4, the tracer concentration of the contributing groundwater can be measured only at times when there is only groundwater flow to the stream, i.e., before and after the runoff events. The temporal variation within the events has to be determined indirectly.

In many reported applications of Eq.(3.3), c_g is given a constant value during the runoff event, equal to the tracer concentration of streamflow just before the event. With this approach, X is often called the fraction of pre-event or pre-storm water (e.g., Fritz et al., 1976). However, Eq.(3.3.) gives the pre-event fraction of water in the stream only if the tracer concentration of all pre-event water that reaches the stream during the runoff event equals the value used on c_g . As discussed in Section 8.1.4, this condition may not be fulfilled for ^{18}O due to the occurrence of pre-event soil water of an isotopic composition different from that of the pre-event groundwater, and also due to spatial variations in the tracer concentration of groundwater.

Another approach is to assume that c_g varies linearly with time between the pre-event and post-event tracer concentrations of streamwater (Hooper and Shoemaker, 1986). With this assumption the fraction X by Eq.(3.3) is larger than the one obtained by use of constant c_g and probably closer to the true groundwater fraction of streamwater.

With the present model (Rodhe, 1984), attempts are made to calculate the time variation of c_g . The model is applicable only for non-reactive tracers, e.g., for stable isotopes but generally not for water chemistry.

Mass balance of water and tracer in the groundwater reservoir in Fig. 3.1 gives,

$$\frac{dV_g}{dt} = Q'_p - Q_g \quad (3.4)$$

$$\frac{d(V_g c_g)}{dt} = Q'_p c_p - Q_g c_g \quad (3.5)$$

where Q'_p is the recharge to the reservoir by new rainwater. With this simple model it is not considered meaningful to distinguish between the soil water and the groundwater reservoirs. They are therefore combined to form one single reservoir, called groundwater reservoir.

If the volume of the groundwater reservoir further is assumed to be constant we get

$$Q'_p = Q_g$$

and, since $Q_g = X Q_s$,

$$dc_g = \frac{Q X}{V_g} (c_p - c_g) dt \quad (3.6)$$

With this equation, the temporal variation of c_g within runoff events can be calculated. The value of V_g is chosen by trial and error so that c_g changes from the pre-event to the post-event value during the event, i.e., between the corresponding values of the tracer concentration of streamwater. It is to be noted that in such applications, Eq.(3.6) is just a crude attempt to describe the time variation of c_g between two known values, using the tracer concentration of the water input and the estimated rate of groundwater flow. With this simple model, V_g should be interpreted as a model parameter rather than a true reservoir volume.

As is pointed out in Section 8.1.4, the variation of c_g during an event cannot be determined accurately, neither from measurements nor by Eq.(3.6). For this reason, the most reliable results are considered to be those obtained when using a constant value of c_g , a procedure that corresponds to the use of an infinite reservoir volume in Eq.(3.6). If c_g changes due to infiltration, such calculations underestimate the fraction of groundwater. Parts of what is called rain- or meltwater [$Q_s(1 - X)$] is then in fact groundwater that has been formed by

newly infiltrated water particles. Bearing this in mind, the term "fraction of groundwater" will be used in the present study, both when c_g is assumed to be constant and when it is assumed to vary according to Eq.(3.6).

The extent of the discharge area is calculated assuming that all rain- or meltwater on such areas reaches the stream as saturation overland flow, while all rain- or meltwater on recharge areas infiltrates (cf. the estimates of stream width by Fritz et al. 1976). The flow of new rain- or meltwater in the stream, Q_p , then equals the water input on discharge areas:

$$Q_p = (P - \Delta S/\Delta t)A Y \quad (3.7)$$

where A is the basin area, ΔS change in snow storage (negative during melting) over the time period Δt , and Y is the discharge area expressed as a fraction of the total basin area.

Since $Q_p = Q_s(1 - X)$,

$$Y = \frac{Q_s}{A (P - \Delta S/\Delta t)} (1 - X) \quad (3.8)$$

Introducing the specific discharge $q = Q_s/A$, we get the basin fraction of discharge area as

$$Y = \frac{q}{P - \Delta S/\Delta t} (1 - X) \quad (3.9)$$

In this study, event mean values of Y are calculated, using event averages of q , X , and $P - \Delta S/\Delta t$.

Eq.(3.9) expresses the area needed for the measured amount of rainfall or snowmelt to generate the observed flow of rain- or meltwater in the stream, as estimated by Eq.(3.3). It is then assumed that overland flow is generated only on discharge areas which are saturated to the surface so that no infiltration can take place. In a more thorough discussion of streamflow generation, presented in Section 10.2, a distinction is made between saturated and unsaturated discharge areas. Eq.(3.9) gives the areal extent of the saturated discharge areas. But since the equation is considered to give a very rough interpretation of the results of the hydrograph separations, the distinction between the two types of discharge areas is not maintained throughout the text. There are of course other possible mechanisms of overland-flow generation, e.g., the generation of

rainfall excess on recharge areas when the rate of water input exceeds the infiltration capacity of the soil. In such cases, the area producing overland flow is larger than the area given by Eq.(3.9), since the water input to one point in the basin may partly infiltrate and partly generate overland flow. Since the extent of saturated areas can be measured by field surveys, Eq.(3.9) is considered to be a valuable means of evaluating the hydrograph separations made by Eq.(3.3). On the other hand, it should be noted that Eq.(3.9) is just a way of interpreting the estimated groundwater fractions in terms of basin characteristics. The groundwater fractions are calculated independently of this interpretation.

3.2 Oxygen-18 as a tracer

In this section the ^{18}O content of precipitation is treated, starting from the basic principles of fractionation. Possible changes of the isotopic composition during water flow through a basin are discussed in Chapter 6.

3.2.1 General occurrence

In natural waters, 0.2% of the water molecules contain the heavy stable oxygen isotope ^{18}O . Different waters show slightly different concentrations of ^{18}O due to a change of isotopic concentration, a so-called fractionation, which takes place during phase transitions. Some fractionation takes place during melting-freezing (see Section 6.2.6), but the most important fractionation governing the occurrence of H_2^{18}O is caused by evaporation-condensation. This latter fractionation can be explained by lower saturation vapour pressure for H_2^{18}O than for H_2^{16}O , which means that the heavy molecules are less inclined to evaporate and more inclined to condense than the lighter ones. During evaporation the remaining liquid is enriched in heavy isotopes and during condensation the remaining vapour is depleted. Strictly speaking, the words evaporation and condensation might be misleading, since fractionation is an effect of the gross molecular fluxes through the interface between liquid and gaseous phases. Gross evaporation is proportional to the saturation vapour pressure at the temperature of the liquid, and gross condensation is proportional to the vapour pressure of the air. For example, enrichment or depletion of liquid water can take place through the exchange of molecules, even if there is no net evaporation.

The ^{18}O -content of a water sample is determined by mass spectrometry. With this method, the sample is compared to a water of known isotopic composition. For this reason, the ^{18}O -content is commonly expressed as the relative deviation of the isotopic ratio of the sample from that of a standard water, generally SMOW (Standard Mean Ocean Water), (Craig, 1961a). The relative deviation is denoted $\delta^{18}\text{O}$.

$$\delta^{18}\text{O} = \frac{R_{\text{SAMPLE}} - R_{\text{SMOW}}}{R_{\text{SMOW}}} \quad (3.10)$$

where the R 's are the isotopic ratios, i.e., the ratios between the number of atoms of the two isotopes, $^{18}\text{O}/^{16}\text{O}$. The δ -value is often expressed in per mil, but in calculations it is more convenient to use fractions directly from Eq.(3.10). In this text the term " ^{18}O -content" will be used in general, qualitative, discussions. When quantitative information is needed and when measured values are presented, the term δ -value or $\delta^{18}\text{O}$ will be used. Per mil (‰) will be used as a unit for ^{18}O -content and its proper meaning as a fraction will be disregarded. Thus, "the δ -value is lowered by 2.1‰" means that

$$\delta_1 - \delta_2 = 2.1\text{‰}, \text{ and not that } \frac{\delta_1 - \delta_2}{\delta_1} = 0.0021.$$

As is common in the literature, $\delta^{18}\text{O}$ is treated as a concentration in many discussions and calculations, in spite of its definition by Eq.(3.10) as a relative deviation in isotope ratio.¹

¹ Eq.(3.3), for instance, also applies to $\delta^{18}\text{O}$ -values, as shown below:

The concentration of ^{18}O , c , and the isotope ratio, R , are

$$c = \frac{^{18}\text{O}}{^{18}\text{O} + ^{16}\text{O}} \quad \text{and} \quad R = \frac{^{18}\text{O}}{^{16}\text{O}} \quad \text{giving} \quad c = R \frac{1}{1 + R}.$$

By using these concentrations in Eq.(3.3) we get

$$X = \frac{R_s - R_p}{R_g - R_s} \quad \frac{1 + R_g}{1 + R_s}$$

and in the δ -notation

$$X = \frac{\delta_s - \delta_p}{\delta_g - \delta_p} \quad \frac{1 + R_{\text{SMOW}}(1 + \delta_g)}{1 + R_{\text{SMOW}}(1 + \delta_s)} = \frac{\delta_s - \delta_p}{\delta_g - \delta_p} G$$

$R_{\text{SMOW}} = 0.00199$ (Craig, 1961a), with the largest conceivable difference between δ_g and δ_s , gives

$$G = 1.0000 + \epsilon, \text{ where } |\epsilon| < 10^{-4}.$$

Isotopic fractionation during phase transition from A to B is described by the fractionation factor α defined by

$$\alpha_{A \rightarrow B} = \frac{R_A}{R_B} \quad (3.11)$$

where R_A and R_B are the isotopic ratios in phase A and B respectively. For the transition from liquid water (subscript l) to vapour (subscript v) α is greater than one and thus $R_l > R_v$.

Eq.(3.11) gives the relationship between the isotopic ratios of vapour and liquid in thermodynamic equilibrium. In this case α is the isotopic equilibrium fractionation factor. Isotopic equilibrium between the phases exists, for example, in a closed container with water and vapour standing for a sufficiently long time for the isotopic ratios to attain constant values. Isotope data on rainwater show that equilibrium conditions prevail during condensation in the atmosphere. During evaporation, however, the fractionation factor is greater than the equilibrium factor, due to a kinetic effect (see further Section 6.1). Throughout this text α denotes the equilibrium isotopic fractionation factor.

The fractionation factor increases with decreasing temperature. For $^{18}\text{O}/^{16}\text{O}$ $\alpha = 1.0098$ at $+20^\circ\text{C}$, 1.0107 at $+10^\circ\text{C}$, and 1.0117 at 0°C (Majoube, 1971).

An alternative, but equivalent, definition of the fractionation factor between liquid and vapour is

$$\alpha = \frac{e_s}{e_{si}} \quad (3.12)$$

where e_s and e_{si} are the saturation vapour pressures for H_2^{16}O and H_2^{18}O respectively. Since fluxes of molecules are proportional to vapour pressures, Eq.(3.12) is suitable when solving problems where rates are concerned, while Eq.(3.11) is suitable when considering phase transitions of amounts, regardless of time.

The different behaviour of H_2^{18}O as compared to H_2^{16}O during evaporation and condensation causes characteristic geographical and temporal patterns in the ^{18}O content of atmospheric water vapour and precipitation. From an analysis of global precipitation data Dansgaard (1964) showed the following so-called "effects" concerning the $\delta^{18}\text{O}$ of precipitation:

The $\delta^{18}\text{O}$ decreases with:

- decreasing surface air temperature, shown as a geographical as well as a seasonal variation
- increasing latitude
- increasing altitude
- increasing distance from the oceans in the directions of vapour transport
- increasing amount of precipitation, other factors being constant

These effects have been fairly well described by simple condensation models (e.g., Dansgaard, 1964): A moist air mass is cooled and the condensate is removed by precipitation, allowing no molecular flux from the liquid back to the vapour (Rayleigh condensation). Since the heavy isotopes are more inclined to condense than the lighter ones, the vapour is gradually depleted of heavy isotopes. At each moment, the condensate is in isotopic equilibrium with the vapour by which it is formed, making the $\delta^{18}\text{O}$ of precipitation gradually decrease as the condensation proceeds. The depletion is accentuated by the increase of the isotope fractionation factor with decreasing temperature. The temperature difference between the vapour supply region on the earth and a certain location is a measure of how far the condensation has proceeded and thus of the depletion in ^{18}O . In the real world this simple model is complicated by, among other things, vapour supply by evapotranspiration during vapour flow outside the original source areas. For this reason, the altitude effect, and in some cases the continental effect, may be well described by the model, whereas the latitude effect needs a more elaborate model (see discussion in Yurtsever and Gat, 1981).

From an analysis of precipitation data from 91 stations around the world, Yurtsever and Gat (1981) concluded that monthly mean surface air temperature was the most important of the effects mentioned above in describing the monthly mean $\delta^{18}\text{O}$ of precipitation on a global scale, with

$$\delta_p^{18}\text{O} = 0.34T - 12.0\text{‰} \quad (3.13)$$

where T is monthly mean temperature in $^{\circ}\text{C}$.

The close relationship found between surface air temperature and precipitation $\delta^{18}\text{O}$ (a correlation coefficient of 0.79 in the regression equation above, for instance) is somewhat surprising,

since the isotope content of precipitation is actually determined by the past and present temperature and isotopic composition at the cloud base of the precipitating air mass. Parallel development of the temperature at the two levels might be one explanatory factor (Dansgaard, 1964).

The effects of isotopic fractionation discussed above are also applicable to the other rare stable isotope in natural waters, deuterium (^2H or D). This isotope occurs as HDO in 0.03% of the water molecules. Encountering the same fractionation processes, but having different α -values, ^{18}O and D in precipitation generally have parallel variations with a known relationship (Craig 1961b).

$$\delta\text{D} = 8\delta^{18}\text{O} + 10 \quad [‰] \quad (3.14)$$

Results from studies of one of the isotopes are thus often valid for the other. This similarity will be used in the present study when referring to the literature, from where reported measurements of δD will be converted to $\delta^{18}\text{O}$ by use of Eq.(3.14).

Due to the kinetic separation effect, which is different for the two isotopes, water that is subjected to evaporation will not follow Eq.(3.14) during enrichment. This deviation, which gives a slope of less than 8 in Eq.(3.14), is a well-exploited tool to identify water that has been subjected to evaporation since the raindrop formation.

3.2.2 $\delta^{18}\text{O}$ of precipitation related to the present study

The isotope used in the present study is ^{18}O . In non-tropical areas there is a prominent seasonal variation in the precipitation $\delta^{18}\text{O}$. In winter the monthly mean δ -values are low as compared to the summer values. Superimposed on the seasonal variation, there is a large short-term variation in the $\delta^{18}\text{O}$ of single precipitation events. These variations are to a large degree reduced in the soil water and even more so in the groundwater reservoir, giving comparatively constant values of groundwater $\delta^{18}\text{O}$, somewhere in between the winter and the summer mean values of precipitation. Therefore, groundwater and precipitation have different δ -values during many precipitation events and often during spring snowmelt. This difference is a prerequisite for hydrograph separation by ^{18}O . The seasonal variation of the precipitation input also makes it possible to estimate transit times and reservoir content for water in a basin.

Monthly samples of precipitation $\delta^{18}\text{O}$ from 16 Swedish stations during the period 1978 to 1985 were analysed by Burgman et al. (1987). Fig.3.2 shows arithmetic means of the monthly δ -values. The locations of the study areas in the present investigation are also shown on the map.

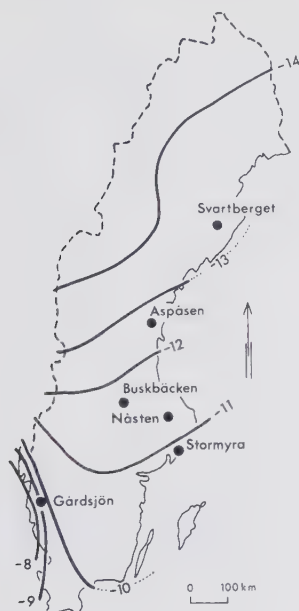


Fig.3.2 Mean $\delta^{18}\text{O}$ of precipitation in Sweden (from Burgman et al. 1987) and location of the present study basins .

As an example of the daily and seasonal variation, precipitation $\delta^{18}\text{O}$ from one of the present study areas, Nästen, is presented in Fig.3.3. The seasonal variation is evident from the monthly values, but the daily values show a large scatter.

Burgman et al. (1987) analysed the seasonal variations at the stations in their network. The amplitudes of the mean seasonal variations at most stations were 2-3‰ or more. The smallest amplitudes, about 1‰, were found at the stations on the western coast of Sweden. Small seasonal variation is obviously a disadvantage when using ^{18}O for studies of runoff processes.

In determining the basin mean of precipitation $\delta^{18}\text{O}$, the altitude effect, i.e., the decrease of the ^{18}O -content with increasing altitude of the site, has to be considered. This effect has been investigated in several studies, often as a means to localize areas of groundwater recharge to large aquifers. The decrease with altitude has generally been found to be $0.1\text{-}0.4\text{‰}(\text{100 m})^{-1}$. The empirical relationship in

Eq.(3.14), with a wet adiabatic lapse rate of $0.6^{\circ}\text{C}(100\text{ m})^{-1}$, gives a decrease in $\delta^{18}\text{O}$ of about $0.3\text{‰}(100\text{ m})^{-1}$.

Due to the comparatively small altitude differences within the investigated basins ($<100\text{ m}$) the altitude effect has been ignored in the present study when determining the $\delta^{18}\text{O}$ of rain-falls. When it comes to the isotopic composition of the snow-pack, however, the altitude effect on precipitation $\delta^{18}\text{O}$, though small in relation to the overall variability of $\delta^{18}\text{O}$, is considered to some extent through basin-wide sampling. (See also Section 7.1.)

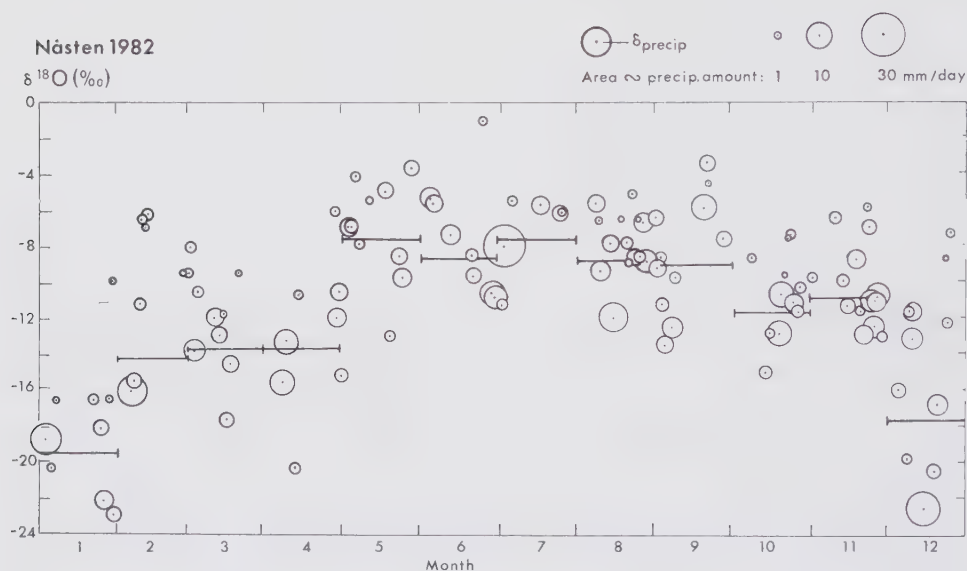


Fig.3.3 Daily and monthly mean values of precipitation $\delta^{18}\text{O}$. Nästen, Uppsala, 1982

In addition to the large scatter of the $\delta^{18}\text{O}$ of daily precipitation samples, shown in Fig.3.3, there may also be a considerable temporal (and spatial) variation in precipitation $\delta^{18}\text{O}$ within individual storms. Three precipitation events of different origin were studied by Ambach et al. (1975). The ranges of $\delta^{18}\text{O}$ within the storms were large, varying between 5 and 10‰ . The δ -value decreased rapidly from the beginning of the storms; it then rose to or above the initial value and finally started to decrease again. The decreases were attributed to the rainout of heavy isotopes from the vapour due to Rayleigh condensation. The rise was assumed to be a consequence of the cell

structure of the precipitation fields, since it could not be explained by any fractionation process. Large in-storm variations of $\delta^{18}\text{O}$ were also reported by Dansgaard (1961) and by Bleeker et al. (1966). Dansgaard further found areal variability due to collector position in relation to the paths of convective showers.

A few examples of the temporal and areal variability of precipitation $\delta^{18}\text{O}$ within single storms from the present investigation are presented in Section 7.1, where the observed relationship between daily δ -values of precipitation and air temperature is also commented upon. The consequences of the temporal variations in precipitation $\delta^{18}\text{O}$ on the hydrograph separations made in this study are discussed in Section 8.2.

4 SITE DESCRIPTIONS

The investigation was performed in 10 basins with areas of 0.03-6.6 km² at 6 localities in Sweden. The basins represent characteristic types of landscape encountered in Sweden, with gneiss or granite rock covered by glacial till soils, vegetated by coniferous forest. The basins were selected to represent various geologic, topographic, and climatic situations. Further criteria in selecting the basins were that they should contain no lakes, that they should be equipped with discharge measuring devices, and that climate data should be available from nearby stations. The possibility of cooperating with other projects in the collection of data was also an important factor.

The basins are named Aspåsen, Buskbäcken, Gårdsjön F1, Gårdsjön F2, Gårdsjön F3, Nåsten, Stormyra and Svartberget (with two sub-basins).

Data on the basins are given in Tables 4.1a, b and c, on the last pages in this chapter.

Svartberget, Aspåsen and Buskbäcken belong to the northern boreal coniferous zone. The climate of these basins has a slight continental influence, with warm summers and comparatively cold winters (Table 4.1c) and a continuous snowpack for 5 months or more. Nåsten is situated at the same latitude as Buskbäcken, but the winters are much milder. Snow cover develops regularly in Nåsten, but considerable melting may occur during winter, due to the lower altitude above sea level and some maritime influence from the Baltic. The climate, and also the day-to-day weather, at Stormyra is rather similar to that at Nåsten. The climate of the Gårdsjön basins, situated 10 km from the North Sea, has a strong maritime influence. The snow storage in these basins is irregular, with melting frequently occurring during the winter. The annual precipitation is larger than in the other basins, a result of the dominating south-westerly winds and a considerable orographic effect.

4.1 Aspåsen (Fig.4.1)

Aspåsen is situated 50 km SW of the city of Sundsvall within the drainage basin of the river Hamrångerån. It is one of the

basins in the area in which the Department of Forest Soils, Swedish University of Agricultural Sciences, investigates the effects on nutrient circulation of increased forest biomass utilization. A few small basins in the surroundings, and the area in general, are described by Rosén (1982). Aspåsen is a first-order basin of simple shape. In its upper parts, the stream flows through a narrow peat strip at the bottom of a distinct valley. The valley flattens downstream and the slope of the ground is directed more in parallel with the stream. The bedrock is covered by sandy to fine-sandy till, which in general is more than one meter thick, with well-developed podzol profiles (Anita Lundmark-Thelin, pers. comm.).

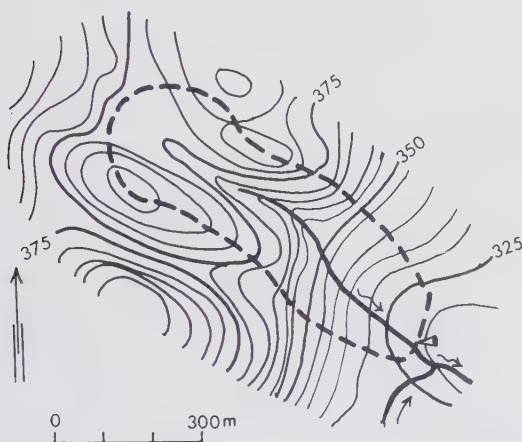


Fig.4.1 Aspåsen basin
(legend, see Fig.4.2).

In 1981, during the first part of the present investigation, the basin was fully forested, mainly by mature spruce. In 1983 the basin was clearcut, except for a 35-year-old stand constituting 14% of the basin area. Thus a supplemental investigation in September 1985 was performed in an almost clearcut basin.

A few preliminary field measurements of saturated hydraulic conductivity gave values ranging from $4 \cdot 10^{-6}$ to $2 \cdot 10^{-4} \text{ ms}^{-1}$ at various depths in the top 0.8 m of the ground, with the highest value at about 0.05 m depth.

In 1981 precipitation was collected in Frisbo, 5 km S of the basin at 160 m above sea level. For the runoff event investigated in 1985, precipitation was collected within the basin at 325 m above sea level.

4.2 Buskbäcken (Fig.4.2)

Buskbäcken is situated in the highland of Bergslagen, 80 km NW of the city of Västerås, within the drainage basin of the river Arbogaån. It is one of the basins in which the "Kloten project" investigated the effects of forestry on water quality and run-off. A general presentation of the area and its hydrology, with special emphasis on the chemistry of streamwater, is given by Grip (1982). Soil, vegetation and waterflow conditions in Masbybäcken, a small basin bordering the Buskbäcken basin, are described in detail by Lundin (1982).

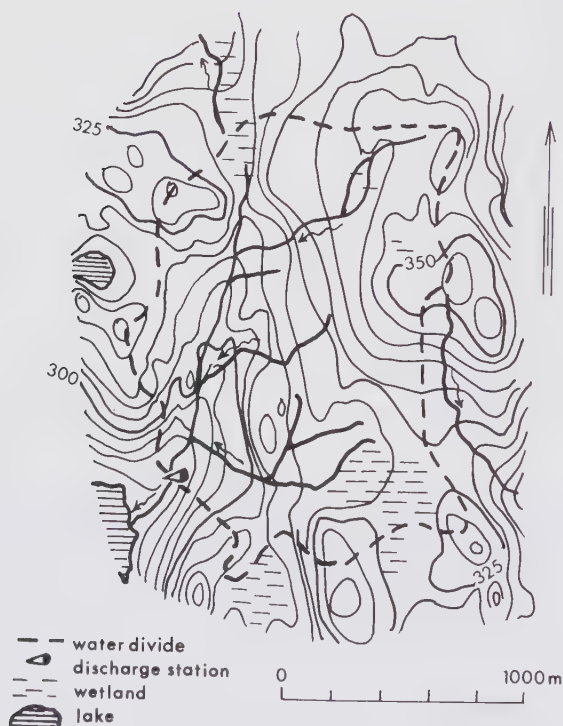


Fig.4.2 Buskbäcken basin.

The forest in Buskbäcken consists of mixed conifers, with small, more or less homogeneous stands of various ages. The drainage network is well developed, with several of the channels dug by man in order to improve forest production. The soil is mostly sandy-silty till. Rock outcrops are rare, existing only at some elevated locations close to the water divide.

Soil¹ depth in most of the basin is probably only a few meters; in the nearby Masbybäcken the maximum soil depth is 3 m. Soil depths >15 m have, however, been observed within and close to the Buskbäcken basin (Rydgren and Sundby, 1982).

The saturated hydraulic conductivity of the soil has been determined in several localities in the nearby Masbybäcken by field and laboratory measurements, showing values around 10^{-4} ms^{-1} at the top few decimetres, decreasing to around 10^{-6} ms^{-1} at 0.5 m depth (Lundin, 1982). There was also a significant decrease in total porosity between these two depths, from 0.50 to around 0.30. These values probably also represent the conditions in the Buskbäcken basin. From a borehole dilution test in one bedrock boring in the Buskbäcken basin, Rydgren and Sundby estimated the saturated hydraulic conductivity of the top 10 m of the bedrock to be about $1 \cdot 10^{-5} \text{ m/s}$.

Buskbäcken, discharge station 61-2227, is a so-called Field Research Basin, operated since 1981 by the Swedish Meteorological and Hydrological Institute (SMHI). Precipitation was collected at Västra Kloten, 4 km S of Buskbäcken discharge station at 280 m above sea level, from which climate station other climate data were also obtained.

4.3 Gårdsjön (Fig.4.3)

The three small Gårdsjön basins are situated 40 km N of Göteborg, 10 km from the sea. They are first-order basins draining into Lake Gårdsjön, which in turn drains into the North Sea by the river Anråseån. The present investigation is part of the multi-disciplinary "Lake Gårdsjön Project", in which the impact of acid deposition on a highly acidified lake and its catchment has been studied. Physiographic and biological features of the area are given by Olsson et al. (1985) and hydrological and geological descriptions by Johansson and Nilsson (1985) and Melkerud (1983) respectively.

The investigated basins are valleys of tectonic origin. In elevated and steep areas the bedrock is mainly bare. In the lower parts of some slopes and in the valleys the bedrock is covered by till, with an estimated mean thickness in basins F1

¹ The term "soil" is used to denote "loose deposits" throughout this report. It is thus used broadly in contradistinction to the bedrock, not in the sense of "the upper part of the loose deposits where soil processes are acting".

and F3 of 2.6 m and 2.1 m respectively (Melkerud, 1983). The streams in the valley bottoms flow mostly through narrow strips of peat soil. A bog area in F2 is located at the upper end of the valley, close to the water divide. The bog contains a very small lake. In F3, on the other hand, a bog is situated in the lowest part of the valley. The basins are forested by mature conifers, except for the eastern part of F3, which was clear-felled in 1976 and later replanted with spruce and pine.

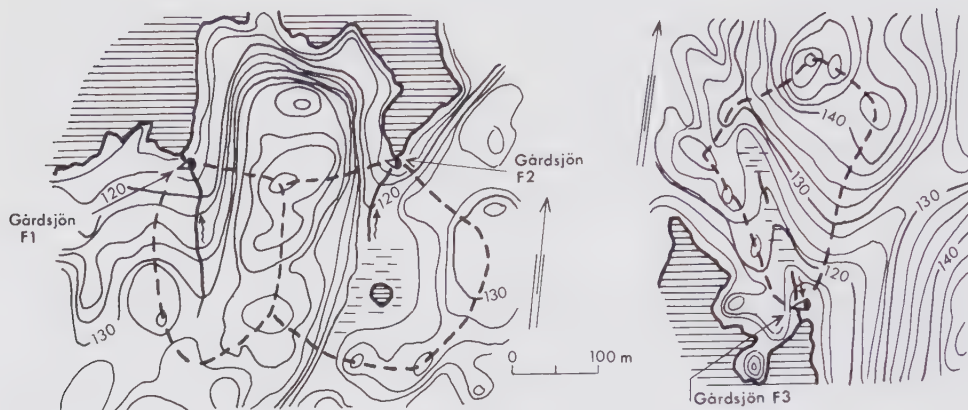


Fig.4.3 The Gårdsjön basins (legend, see Fig.4.2).

Melkerud (1983) classified the till in the basins as fine sandy-silty, with 5-10% clay and a low gravel content. He analysed several samples from F1 and F3 and found very little variation in pore-size distribution. Field measurements of saturated hydraulic conductivity in the top 0.5 m of the soil at several localities around Lake Gårdsjön generally yielded values between $1 \cdot 10^{-5}$ and $5 \cdot 10^{-4} \text{ ms}^{-1}$, with the highest values occurring close to ground surface (G. Lindström, pers. comm.).

Precipitation was sampled 0.8 km NW of the basin F3, at Lilla Komperöd, 115 m above sea level, where air temperature was also measured. During two intensively sampled runoff events in June and July 1982 (named events A and B), precipitation was sampled 0.5 km W of basin F1, at 125 m above sea level.

4.4 Nåsten (Fig.4.4)

The Nåsten basin is situated 4 km SW of central Uppsala. It drains by the river Uppsala-Näsbäcken to Lake Ekoln, a bay of Lake Mälaren. The physiography and hydrology of the basin are

described by Bergqvist (1971a, b). It is an area of low relief, characterized by exposed bedrock and a thin sandy-silty till cover. In the depressions the till is frequently covered by glacio-lacustrine clay. The clay areas are, or have been, cultivated. Considerable flooding occurs in the valley bottoms at times of high discharge, but in its lowest reaches the stream channel is deep and flooding does not occur. The drainage network is in many places influenced by man, as reflected by several ditches draining the abundant bog areas and by covered drains in the agricultural fields. At a nearby site, saturated hydraulic conductivities of around $3 \cdot 10^{-6} \text{ ms}^{-1}$ were observed in the top 0.2 m in the till soil. The forest is composed of mixed conifers. Considerable portions of the forested areas have been clearcut and are now vegetated by young stands.



Fig.4.4 Nästen basin
(legend, see Fig.4.2).

The discharge station, 61-1742 Stabby, is operated by SMHI. Precipitation was collected 200 m E of the discharge station, at about 25 m above sea level. Climate data were obtained from Ultuna Climate Station, Swedish University of Agricultural Sciences, 4 km NE of the basin outlet.

4.5 Stormyra (Fig.4.5)

Stormyra is situated in Tyresö, 20 km SE of central Stockholm. The stream drains into the Baltic 300 m downstream of the basin's outlet. Physiographical and hydrological data on the basin are given by Liljeqvist (1973). Like Nåsten, this basin is characterized by exposed bedrock, thin till deposits and in the valley floors glacio-lacustrine clay. The relief, however, is more pronounced, reminiscent of that at the Gårdsjön basins. Soil of any significant thickness is found only in the lower parts of some slopes and in the valley bottoms. Gotthardsson (1973) reports that 50% of the basin has a soil thickness of less than 10 cm. A flat area upstream from the basin outlet is meadow-land, with some slightly elevated parts vegetated by deciduous forest. At periods of high discharge, large parts of the meadow area are flooded. The remainder of the basin is vegetated by mixed conifers, mostly dominated by pine, but with spruce dominating in many low-lying areas. The observed runoff from Stormyra is considerably larger than the one from Nåsten, although the climate is similar in the two basins (Table 4.1c). The difference is probably a result of the very thin soil cover in Stormyra, giving little soil water storage and thus small evapotranspiration.

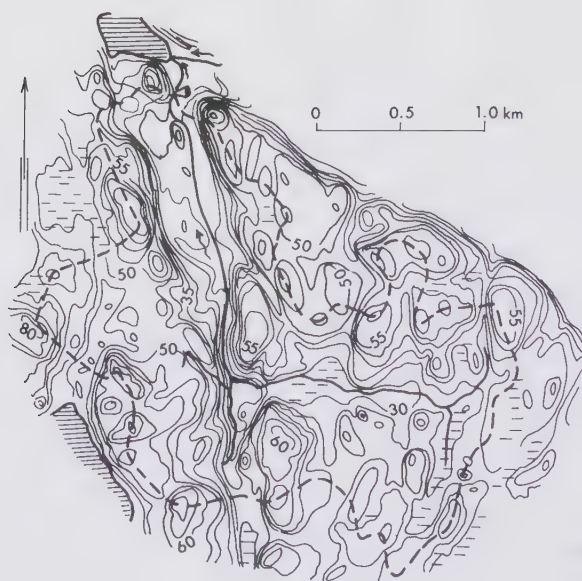


Fig.4.5 Stormyra basin
(legend, see Fig.4.2).

Stormyra, with discharge station 62/63-1835, is a Field Research Basin operated by SMHI. During the first year of the present study, precipitation was collected at Stormyra Climate Station, 300 m NE of the basin outlet, at 10 m above sea level. In June 1980, the precipitation gauge was moved to a place 100 m E of the outlet at 20 m above sea level. Other climate data were obtained from the original location throughout the study.

4.6 Svartberget (Fig.4.6)

Svartberget is situated 60 km NW of Umeå. The basin is a part of Vindeln Forest Research Station of the Swedish University of Agricultural Sciences. The geology and vegetation of the experimental area are described by Tamm (1925) and Malmström (1925) respectively. A detailed geological map of the basin is given in Aastrup et al. (1982) and some aspects of the hydrology are discussed by Sokollek (1985).

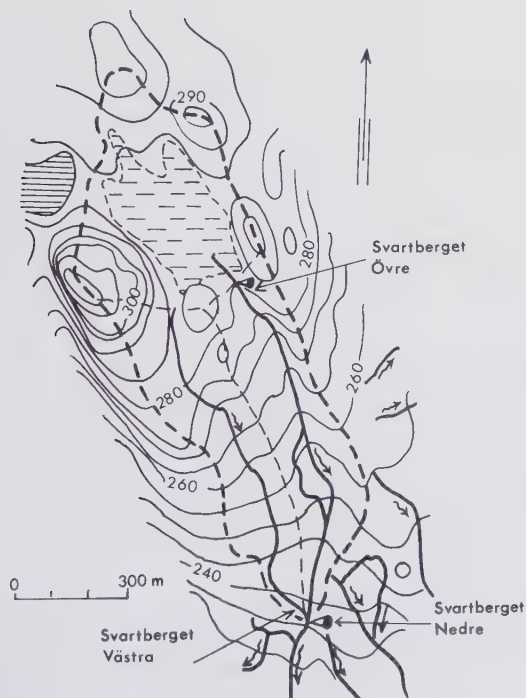


Fig.4.6 Svartberget basin (legend, see Fig.4.2).

Within the total basin, discharged through Svartberget Nedre (=lower), two sub-basins of very different physiography have

been investigated. One, Svartberget Övre (=upper) is dominated by a bog, Stortjärnmyren, constituting 40% of the area of this sub-basin. The other, Svartberget Västra (=west), on the other hand, contains almost no bog area.

The main stream in its downstream reach, and the stream in Svartberget Västra sub-basin, flow partly through peat soils developed on the valley bottoms. The extent of these peat areas gradually increases downstream towards the basin outlet.

The basin is forested by mature pine and spruce. Pine predominates on the ridges and other elevated areas and spruce in low-lying areas. The central part of Stortjärnmyren has no trees while the outer part is sparsely vegetated by pine trees of low productivity.

The basin is covered by glacial till, probably several metres thick in general. Lenses of coarse, sorted sediments within the till have been found in several places (H. Ivarsson, pers. comm). There is only one small rock outcrop, at a little hill just south of Stortjärnmyren. The presence of the highest post-glacial coast line in the basin, at 255-260 m above sea level (Tamm, 1925), is reflected by occurrences of littoral gravel at this level. Preliminary measurements of saturated hydraulic conductivity gave values from 10^{-4} to 10^{-7} ms^{-1} in the top 0.8 m of the mineral soil (H. Grip, pers. comm.). The values were very scattered, but conductivities below 10^{-6} ms^{-1} were found only below 0.5 m depth. Below this level, lenses of coarse sorted material are found, which may greatly affect the subsurface water flow.

The stream channels were dug by man for drainage purposes. Thus, the main channel has almost vertical sides and the water level generally lies about 1 m below the surrounding soil surface. For this reason, saturated areas developing at the foot of slopes may only occasionally reach the stream, which in many reaches is surrounded by unsaturated peat soils.

The observed mean runoff at Svartberget Nedre (252 mm year^{-1}) is considerably smaller than the value given in Table 4.1c (320 mm year^{-1}), which refers to the 661 km^2 Sävarån basin, discharge station Stenfors, situated 25 km E of Svartberget. The available data cover different time periods, Nov.1981-Oct.1985 at Svartberget and 1951-1979 at Stenfors. But since the mean precipitation at the nearby Hällnäs-Lund is about the same for the two periods, 587 as compared to 567 mm year^{-1} , the

mean runoff should be similar for the two basins. The smaller runoff observed at Svartberget suggests a considerable groundwater leakage from this basin. Svartberget basin is part of a landscape sloping gently (by about 3-5%) in a SE direction. The water divide is well accentuated along the NE and SW sides of the basin, defined by ridges parallel to the stream. In the lower part of the basin, however, there may be a groundwater leakage. The coarse layers observed in the soil may play an important role in such a leakage.

The discharge station at Svartberget Nedre was working throughout the study, but the one at Svartberget Övre was not working properly. The discharge at Svartberget Västra was not measured. Precipitation was collected at the field laboratory of Vindeln Forest Research Station, 1 km SW of the basin outlet, at 225 m above sea level.

Table 4.1a Site descriptions: Physiography.

Basin	Area (km ²)	Altitude			Mean slope ¹			Lat. N	Long. E
		min (m a.s.l.)	max	max-min (m)	basin S _{b1} (%)	basin S _{b2} (%)	stream S _s (%)		
Aspåsen	0.17	320	390	70	16	17	9	62° 01'	16° 42'
Buskbäcken	1.84	280	354	74	7	6	3	59° 54'	15° 49'
Gårdsjön	F1 0.036	115	135	20	14	11	10	58° 03'	12° 02'
	F2 0.040	115	135	20	15	10	5	58° 03'	12° 01'
	F3 0.028	115	150	35	18	21	8	58° 04'	12° 01'
Nåsten	6.6	18	55	37	4	1	0.3	59° 48'	17° 35'
Stormyra	4.0	10	77	67	10	4	0.3	59° 12'	18° 20'
Svartberget									
Övre	0.20	285	310	25	7	6	<2	64° 16'	19° 46'
Västra	0.16	235	310	75		19	5	64° 15'	19° 46'
Nedre	0.50	235	310	75	7	11	5	64° 15'	19° 46'

¹ S_{b1} = mean slope calculated by the method of Wentworth (1930). See below.
 S_{b2} = $\Delta h / \sqrt{A}$ where A = basin area and Δh = max. altitude difference,
 S_s = mean slope of the main stream from the centre of the basin to the outlet (approximate figures).

Calculation of mean slope by the method of Wentworth (1930): Slope at a certain point is here defined as $\tan \alpha$ (given as percentage), where α is the angle between the ground plane and the horizontal plane. The mean slope of an area in a certain direction is determined in the following way: A system of parallel lines, separated by equal lengths, is put on a topographic map with the lines oriented in the desired direction. The number of intersections with the contours, n , is counted. The mean distance between the contours is given by $(L/n) \cdot (2/\pi)$, where L is the total length of the parallel lines. The mean slope, S , in the direction under consideration is

$$S = \frac{\pi}{2} \cdot \frac{ne}{L}$$

where e is the altitude difference between the contours of the map. The thus calculated mean slopes vary in different directions, reflecting the orientation of the fault systems and the aspects of the valleys. The mean slopes given in Table 4.1a are arithmetic means of mean slopes calculated in four directions, N-S, NE-SW, E-W and SE-NW.

A more easily obtained measure of slope is also given in the table, namely $\Delta h / \sqrt{A}$. The two values agree qualitatively. The slope in the largest basins is, however, underestimated by the simple method since it does not account for the broken relief seen at this scale.

Table 4.1b Site descriptions: Geology and land use.

Basin	Out-crop ¹	Soil cover (%)			Post-glacial coast line ³	Domi-nating bed-rock	Land use (%)		
		Till ²	Clay	Peat			Forest ⁴	Clearing	Field or meadow
Aspåsen	<2	>98	0	0	A	granite	100	0 ^a	0
Buskbäcken	<5	>86	0	9	A	granite	80	19	1
Gårdsjön F1	47	53	0	0	A,B	gneiss	93	7	0
F2	47	40	0	13 ^b	A,B	and	99	0	0
F3	32	63	0	5	A,B	granite	47	53	0
Nåsten	25	55	16	4	B	granite	87		13
Stormyra	67	13	12	8	B	gneiss, granite	92		8
Svartberget									
Övre	0	60	0	40	A	gneiss	90	10	0
Västra	0	80	0	20	A,B	gneiss	100	0	0
Nedre	0	67	0	33	A,B	gneiss	96	4	0

The figures on geology and land use for the three largest basins, Nåsten, Stormyra and Buskbäcken, have been obtained from geological and topographical maps of scale 1:50 000 and areal photographs. Data on the other basins are from various special investigations (see references given in the text).

- ¹ Outcrops are defined as areas with a soil cover <0.5 m except for in the Gårdsjön basins where such areas are defined by a soil cover <0.2-0.3 m.
- ² In Svartberget Västra and Svartberget Nedre the figure includes occurrences of littoral gravel (5% and 2% of the basin areas respectively).
- ³ Altitude of the basin in relation to the highest post-glacial coastline: A = above, B = below.
- ⁴ Including open bog areas, of significance particularly in Svartberget Övre and Nedre, see text.
- ^a At the first part of the investigation.
- ^b Including a lake, 1% of the basin area.

Table 4.1c Site descriptions: Climate and water balance.

Basin	Air temperature (°C) ¹			Water balance (mm·year ⁻¹) ²		
	Year	July	Jan.	Precip. ³	Runoff	Evapotransp. ⁴
Aspåsen ^a	2.3	14.5	-8.8	755 ^b	330 ^c	430
Buskbäcken ^d	5.0	16.5	-6.0	890 ^e	435 ^f	460
Gårdsjön ^g	6.5	16.6	-2.8	1140 ^h	606 ⁱ	530
Nåsten ^j	5.4	16.3	-4.4	660	204 ^k	460
Stormyra ^l	5.6	17.5	-2.6	695 ^m	303 ⁿ	390
Svartberget ^o	1.0	14.5	-12.1	700	320 ^p	380

1 July is the warmest month in all basins and January is the coldest in the three northernmost basins. In Nåsten, Stormyra and Gårdsjön, February is colder than January by about 0.5°C in the monthly means.

2 The figures given refer to periods of different lengths and the periods for various elements of the water balance in a certain area do not always coincide. In some basins the observation periods are much too short to represent long-term annual means.

3 With the exception of those for Gårdsjön, the figures on precipitation are normal values for the period corrected for measurement errors given by Eriksson (1983). In these cases the applied correction increased the observed yearly amounts by 15-25%.

4 Evapotranspiration is calculated as the difference between the figures given on precipitation and runoff, put in round figures.

^a Air temperature from Delsbo, 75 m a.s.l., 25 km SW of Aspåsen, 1931-1960, corrected for altitude difference by -0.6°C (100 m)⁻¹.

^b Strömbacka, 165 m a.s.l., 8 km S of Aspåsen.

^c Franshammar, Rexforsån, 1931-1960.

^d Air temperature 1931-1960 according to Grip (1982).

^e Grängesberg, 25 km NW of Buskbäcken.

^f Buskbäcken 1970-1984.

^g Air temperature from Ljungskile, 70 m a.s.l., 20 km N of Gårdsjön, corrected for altitude difference by -0.6°C (100 m)⁻¹.

^h Oct. 1979 - Sept. 1984 (T. Nilsson, pers. comm.), corrected by +17%.

ⁱ Runoff from the whole Lake Gårdsjön basin (2.1 km²) Oct. 1979 - Sept. 1984 (T. Nilsson, pers. comm.).

^j Air temperature and precipitation from Ultuna 1951-1980.

^k Nåsten (Stabby) 1963-1983.

^l Air temperature from Gustavsberg, 10 m a.s.l., 15 km N of Stormyra, 1931-1960.

^m Tyresö, 5 m a.s.l., 5 km N of Stormyra.

ⁿ Stormyra 1963-1983.

^o Air temperature and precipitation from Hällnäs-Lund, 181 m a.s.l., 7 km E of Svartberget, 1951-1980.

^P Stenfors, Sävarån 1951-1979.

5 MEASUREMENTS

The main purpose of this study is to estimate the relative contributions to streamflow made by new rain- or meltwater and groundwater. The fraction of groundwater (or pre-event water) is calculated by Eq.(3.3) using ^{18}O as a tracer. For this purpose, regular sampling for ^{18}O was performed in precipitation, snowpacks and streamwater. Groundwater was also sampled, but more irregularly, in order to get indirect information on the $\delta^{18}\text{O}$ of groundwater contributing to streamflow during events, a variable which cannot be measured directly.

In some methodological discussions the $\delta^{18}\text{O}$ data are supplemented with data on the electrical conductivity of the water, which was measured on most samples taken for ^{18}O .

Sampling for ^{18}O in some of the basins started in 1979, and the results presented in this report cover the period up to 1982 except for a few later measurements. Within these years, the sampling was performed for various periods and with various intensities in the different basins, as shown by the chart in Fig.5.1.

Basin	1979	1980	1981	1982
Aspåsen			— (+Sept. 1985)	
Buskbäcken				
Gårdsjön F1				
F2				
F3				
Nåsten	—	—		—
Stormyra				
Svartberget Övre			—	
Nedre			—	
Västra				

— Mainly daily sampling
 Weekly to monthly sampling

Fig.5.1 Measurement periods for the $\delta^{18}\text{O}$ of streamwater.

5.1 Field measurements

5.1.1 Precipitation

Precipitation was normally collected in the standard gauge of the Swedish Meteorological and Hydrological Institute (SMHI). The gauge is cylindrical with an opening of 200 cm². For rain-water collection it is equipped with a funnel to prevent evaporation. The cylinder is surrounded by a wind shelter to reduce aerodynamic losses. In the normal operation, as in this study, the gauge orifice is 1.5 m above the ground. For snowfall collection at some of the stations, cylinders having similar collection dimensions were placed about 0.7 m above the ground.

According to Eriksson (1983) the SMHI gauges may underestimate the yearly precipitation by 15-30%, mainly due to aerodynamic losses but also due to evaporation. The aerodynamic losses are particularly large for snowfall. They may be greatly reduced if the gauge is placed at a well-sheltered location. The gauges used in the present study are comparatively well located with regard to aerodynamic losses. With the exception of the figures on long-term mean values given in Table 4.1c, no correction has been applied to the collected amounts. Aerodynamic losses introduce little error in the determination of the $\delta^{18}\text{O}$ of single rainfalls or snowfalls, since a certain fraction of the precipitation is collected at each moment.

Saxena (1987) found that water stored in SMHI gauges could be slightly enriched in ^{18}O by molecular exchange before sampling. The observed enrichment was about 0.2‰ day⁻¹ of 10 mm stored water and 0.06‰ day⁻¹ of 30 mm stored water. In order to diminish ^{18}O enrichment in the gauges, the holes in the funnels were reduced from about 100 to about 10 mm². With this modification of the SMHI gauge, enrichment can be disregarded in daily sampling.

The gauges were normally sampled daily. For certain periods in some of the basins, rainfall was sampled twice a day. During the rainfall events in the Gårdsjön basins in June and July 1982 intensive rainfall sampling was undertaken, with several samples a day. Water samples from the collected snowfalls were taken by putting an air-tight cover on the cylinder and bringing it indoors to allow the snow to melt.

Precipitation collected at one station, within or close to the basins, was assumed to be representative of the precipitation

input to the basins. The precipitation stations were located in the lower parts of the basins or, when outside of the basins, at similar or lower altitudes (see Chapter 4). The altitude difference between the mean altitude of the basin and the precipitation station was comparatively large in Aspåsen during 1981, about 200 m. In the other basins, including Aspåsen during 1985, it ranged from about 65 m (Svartberget Övre) to about 5 m (Gårdsjön F3). Due to the systematic decrease in the $\delta^{18}\text{O}$ of precipitation with altitude, about $0.2\text{--}0.4\text{‰}(100\text{ m})^{-1}$ (cf. Section 3.2.2), the gauge at Aspåsen might have overestimated precipitation $\delta^{18}\text{O}$ by some tenths of ‰ during 1981, whereas the altitude effect is negligible for all other cases. However, the precipitation collected at Aspåsen during 1981 was only used for judging the precipitation input to the snowpack (cf. Section 7.2). During the only rainfall-generated runoff event investigated in this basin, rainwater was collected within the basin.

The distance between the precipitation station and the most remote part of the basin ranged from 5.3 km (Buskbäcken) to 0.5 km (Gårdsjön F1, summer events 1982). With these distances, accurate basin mean values of precipitation amount and $\delta^{18}\text{O}$ were probably obtained for rain of frontal origin, but for convective showers less representative samples may have been obtained in some instances.

5.1.2 Snowpack

Except for in a few preliminary experiments with snow lysimeters, the isotope content of meltwater was not measured directly in this study. The $\delta^{18}\text{O}$ of meltwater was assumed to be equal to that of snowpack, as discussed in Section 7.2. In order to examine the changes in snowpack $\delta^{18}\text{O}$ during winter and spring, the samplings of the total snowpack were complemented by less intensive profile samplings.

The $\delta^{18}\text{O}$ of snowpack was determined from vertical snow cores, taken with cylindric snow tubes with an inner diameter of 4.0 cm. When sampling small to moderate snowpacks (water equivalent less than about 100 mm) 2–4 snow cores were taken at each station and the snow or melted water was mixed to form one composite sample from the station. To ensure a random sampling distribution, the cores were taken in a predetermined geometric configuration. For practical reasons, only one snow core per station was sampled for ^{18}O analysis from deep snowpacks, but

the weight and depth of the snow cores were generally observed at 4 points at each station.

Snow samples from snowpacks were melted at room temperature in air-tight plastic bags. After complete melting the samples were stored in glass bottles.

The snowpack sampling stations were distributed over the basins to give basin mean values, either equidistant along straight lines or at selected sites accessible from roads.

In Aspåsen, Buskbäcken, Stormyra during 1980 and 1982 and in Svartberget Nedre, 10 stations were sampled on each survey date. In Nåsten and Stormyra during 1979, 3-6 stations were used, and in Nåsten during 1980 and 1982, 6 stations. In Gårdsjön F1 and F3 during 1980, 5 stations were sampled. The number of usable stations decreased at the end of the melt periods as some of the stations became snowfree. Basin mean values from Svartberget Nedre were also assumed to represent the two sub-basins, Svartberget Övre and Svartberget Västra, although only a few of the sampling stations were actually situated inside these sub-basins.

During melting periods, the snowpack was generally sampled once a week, but intenser (Buskbäcken and Gårdsjön F1 and F3 during 1980) as well as sparser (Buskbäcken during 1981 and 1982 and Stormyra) sampling occurred. Before melting, the snowpacks were sampled about once a month.

The vertical ^{18}O layering of the snowpacks was investigated by using 20-cm-long samplers with rectangular dimensions of 5 cm by 7 cm. The samplers were inserted horizontally into the wall of a pit dug immediately before the sampling. In more shallow snowpacks, the profiles were sampled and analysed for ^{18}O in 5-cm layers. In deeper snowpacks, adjacent samples were mixed to form composite 10-cm samples. On each date that profile samples were taken, a new pit was dug a few metres from the earlier ones, at any given sampling location.

Vertical snow profiles were taken in some of the basins during the accumulation period and in all basins during the spring floods. The most complete profile sampling was undertaken in Buskbäcken during 1980 (2 stations, twice a week), Nåsten during 1982 (1 station, weekly), and in Svartberget during 1981 and 1982 (2 stations, weekly).

5.1.3 Groundwater

Groundwater was sampled at irregular intervals from springs and wells. Naturally flowing springs were present in the Buskbäcken, Nåsten, and Svartberget basins. Well samples were obtained from household wells in some basins and from observation wells in Buskbäcken and in the Gårdsjön basins. The former are boreholes in the bedrock, about 20 m deep. The wells in the Gårdsjön basins are plastic tubes installed in the loose deposits, with intakes about 0.4-0.9 m below the ground surface. Before sampling these tubes, they were emptied by pumping and allowed to refill.

5.1.4 Streamwater

In all basins, discharge was measured by triangular weirs fitted with recording gauges. Streamwater was sampled manually, mostly at the weirs of the discharge stations. By sampling from the weirs' outflow, a good mixing of the streamwater was assured and simultaneous values of stream discharge were obtained. For a few periods of very low discharge, however, isotope enrichment and also damping of tracer fluctuations in the stilling pond may have influenced the results. In order to avoid this problem, streamwater was sampled at the pond inlets during some of the intensely sampled rainfall events.

During snowmelt, streamwater was normally sampled once a day, close to the hour of daily maximum flow, which occurred in the evenings. The spring floods of Buskbäcken 1980 and Gårdsjön F3 1980 were intensively sampled, with about four samples a day, and those of Buskbäcken 1981 and Gårdsjön F1 and F2 1980 were sampled twice a day. The spring floods sampled in 1981 and 1982 were intensively sampled for a few days close to the day of spring flood culmination.

During the remainder of the year, streamwater was normally sampled about once a week, except for in Buskbäcken 1980 and 1981 where samples were taken only once a month. Daily or more frequent sampling was performed during selected rainfall-generated runoff events with attempts being made to obtain the greatest sampling frequency during the hydrograph rise and peak-flow. The most intensive sampling was performed in Gårdsjön F1 in June 1982, where up to 13 samples were taken per day.

5.2 Laboratory measurements

Oxygen-18

The relative $^{18}\text{O}/^{16}\text{O}$ ratios of the water samples were measured with a Varian-mat mass spectrometer. The instrument measures the mass of CO_2 molecules in a CO_2 gas in equilibrium with the liquid water sample. The obtained $^{18}\text{O}/^{16}\text{O}$ ratio of the sample is compared to that of a reference water with known $\delta^{18}\text{O}$ -value as compared to SMOW. The accuracy of the $\delta^{18}\text{O}$ -values of the samples is $\pm 0.1\text{--}0.2\text{‰}$.

Electrical conductivity

The electrical conductivity of the water samples, γ , was measured at $+25^\circ\text{C}$. According to the manufacturer the accuracy of the instrument is about $\pm 1\%$. The pH-value of the samples was also measured, but since the measurements were generally made several days or weeks after the sampling, these pH-values may differ considerably from those in the field. Therefore the pH-values were used only to eliminate the effect of hydrogen ions on the conductivity so that the conductivity was solely a measure of the concentration of dissolved solids. The following correction was used (cf. Handbook of Chemistry and Physics)

$$\gamma_{\text{corr}} = \gamma - 3500 \cdot 10^{-\text{pH}} \quad (\text{mS} \cdot \text{m}^{-1})$$

Thus, at pH-values of 6, 5, and 4 the corrections are -0.035 , -0.35 and $-3.5 \text{ mS} \cdot \text{m}^{-1}$ respectively. Considering the accuracy of the conductivity measurements, the correction becomes important at $\text{pH} < 5.5$.

6 FRACTIONATION PROCESSES FOR OXYGEN-18 WITHIN A BASIN

The use of ^{18}O as a tracer in hydrology requires a knowledge of the possible rates of fractionation during water flow through a basin. In the present study the δ -values of precipitation or snowpack are measured to give the ^{18}O input to the basins and those of streamwater to give the output. In order to interpret the δ -values in terms of hydrological processes, any changes in the δ -value other than by the mixing of different waters, i.e., any changes by fractionation, must be taken into consideration and either estimated or treated as errors.

Fractionation within a basin may occur by evaporation and molecular exchange with atmospheric vapour in intercepted water, soil water, snowpacks and in surface water bodies. Fractionation also takes place between snow and meltwater during snow-melt. These and a few other possible fractionating processes are discussed in this chapter in order to give a background to the use of ^{18}O for hydrograph separation and for modelling the flow of water through a basin. Attempts are made to quantify the isotopic changes during various stages of the water flow. This is done by applying a model for isotopic change in a theoretical water body, which is developed in the first part of the chapter.

6.1 Models for fractionation due to evaporation and exchange with water vapour in the atmosphere

The basic principles for fractionation during phase transition were discussed in Section 3.2.1. For water bodies exposed to the atmosphere, both the kinetic effect and the effect of mass exchange with the atmospheric water vapour must be considered. Once the isotopic flux through the surface is known, isotope and water balances for the water body can be established to give the change in $\delta^{18}\text{O}$ of the water with time.

6.1.1 Models for the $\delta^{18}\text{O}$ of the net evaporate

Models for fractionation during evaporation into a moist atmosphere were discussed by Craig and Gordon (1965). In one of their models, which has come to form the basis for many subsequent works, the molecular flux from the liquid surface to

the measurement level in the atmosphere is supposed to take place through a thin laminar and a turbulent layer, characterized by certain resistance values. Isotopic equilibrium is assumed to exist between liquid and vapour at the surface. Since the molecular diffusion coefficients for vapour flow through the air, which determine the flow rates within the laminar sublayer, are somewhat smaller for $H_2^{18}O$ than for $H_2^{16}O$, an additional fractionation takes place just outside the liquid surface. The turbulent fluxes, on the other hand, are assumed to be non-fractionating, carrying all isotopes with the same efficiency.

From the above considerations Craig and Gordon deduced a formula for the isotopic composition of the net evaporate. A similar formula can be deduced in a simple way as follows:

The net water flux through a surface is proportional to the difference in vapour concentration at the surface and in the atmosphere, which in turn is proportional to the difference in vapour pressure. Similarly, the net flux of heavy isotopes (subscript i) is proportional to the difference in isotope concentration, which is given by the difference in vapour pressure times the isotope ratio. Following the approach by Eriksson (1961, 1965) and remembering that $e_{si} = e_s/\alpha$ (Eq.3.12), we get the mean net fluxes of heavy and abundant isotopes as

$$E_i = \frac{1}{r_i} \left(e_s \frac{1}{\alpha} R_s - e_a R_a \right) C \quad (6.1)$$

$$E = \frac{1}{r} (e_s - e_a) C \quad (6.2)$$

where E = net evaporation

r = total resistance from the surface to the measurement level in the atmosphere

e_s = vapour pressure at the surface (= saturation vapour pressure at the surface temperature)

e_a = vapour pressure in the atmosphere

R_s = isotopic ratio in the liquid

R_a = isotopic ratio in the atmosphere

C = a factor relating specific humidity to vapour pressure ($C \approx (M_v/M_a) \cdot (1/p)$ where M_v and M_a are the molecular weights of water vapour and dry air, respectively, and p is air pressure)

C/r = the so-called transfer coefficient or wind function, $f(u)$, appearing in aerodynamic evaporation formulae.

α = equilibrium isotopic fractionation factor

The isotopic ratio of the net evaporate, R_E , is then

$$R_E = \frac{E_i}{E} = \frac{r}{r_i} \frac{(e_s \frac{1}{\alpha} R_s - e_a R_a)}{(e_s - e_a)} \quad (6.3)$$

Using the δ -notation (Eq. 3.10) and introducing a kinetic separation factor defined by

$$\beta = \frac{r_i}{r}$$

Eq.(6.3) becomes

$$\delta_E = \frac{\frac{e_s}{\alpha} (\delta_s + 1) - e_a (\delta_a + 1) - \beta (e_s - e_a)}{(e_s - e_a) \beta} \quad (6.4)$$

The vapour pressures appearing in Eq.(6.4) can be eliminated by means of the "vapour-pressure ratio" (often called the normalized relative humidity) defined as

$$h = \frac{e_a}{e_s}$$

giving

$$\delta_E = \frac{\frac{1}{\alpha} (\delta_s + 1) - h (\delta_a + 1) - \beta (1 - h)}{(1 - h) \beta} \quad (6.5)$$

Eq.(6.5) is similar to the widely used Craig and Gordon Formula

$$\delta_E = \frac{\frac{1}{\alpha} (\delta_s + 1) - h (\delta_a + 1) - 1 - \Delta\epsilon}{1 - h - \Delta\epsilon} \quad (6.6)$$

where $\Delta\epsilon$ = a kinetic separation factor defined by Craig and Gordon as $\Delta\epsilon = (1 - h)(r_i/r - 1)$.

Since $\Delta\epsilon$ is dependent upon h it is considered more convenient to use β as a kinetic separation factor. The meaning of β is readily seen in the case of evaporation into a dry atmosphere (Rayleigh evaporation). Eq.(6.3) then reduces to

$$R_E = \frac{1}{\alpha \beta} R_s \quad (6.7)$$

The effective fractionation factor in this case is thus $\alpha \beta$.

In situations when the air temperature is equal to the surface temperature, Eq.(6.5) is most useful since the vapour-pressure ratio is equal to the relative humidity of the air. When the two temperatures differ, as in the case of snowmelt, Eq.(6.4) is preferable.

The process of non-equilibrium fractionation, as reflected by r_i/r or β , is not yet fully understood. Various values of r_i/r have been proposed, the general approach being

$$\frac{r_i}{r} = \left(\frac{D}{D_i} \right)^n$$

where D_i and D are the molecular diffusion coefficients for vapour flow through air and n is a parameter depending on the flow mechanism, often having a value around 0.5. Commonly assumed values of $\Delta\epsilon/(1-h)$ are 0.013-0.016 (Gat 1970), equivalent to $\beta = 1.013-1.016$.¹

In the deduction of the expression for δ_E , only kinetic effects caused by different rates of molecular diffusion for the isotopic species in the laminary atmospheric sublayer have been taken into account. Additional kinetic fractionation effects may occur at the surface or in a laminar sublayer of the liquid. This latter effect was included in the original formula by Craig and Gordon, but in applications of the formula it is generally disregarded. In laboratory experiments with falling

¹ The molecular diffusion coefficients for water vapour in gases have been determined experimentally by Merlivat (1978) as

$$\frac{D(\text{H}_2^{16}\text{O})}{D(\text{H}_2^{18}\text{O})} = 1.029.$$

With $n = 0.5$ this value gives $\beta = 1.014$.

raindrops, Stewart (1975) found that kinetic fractionation took place solely at the atmospheric laminar sublayer. No kinetic fractionation was found at the surface or within the liquid, and he concluded that a surface kinetic effect would never be of significance in nature, but that kinetic effects within the liquid could not be excluded in other situations.

Pan experiments by Allison and Leaney (1982) showed that δ_E from constant volume evaporation pans could be well predicted by Eq.(6.6) with $\Delta\epsilon/(1-h) = 0.016$, i.e., $\beta = 1.016$.

6.1.2 Model for isotopic change in a well-mixed water body

Given the isotopic composition of the net evaporate, equations for isotopic change within a well-mixed water body (Fig.6.1) is deduced from water and isotope balances of the water body. The equations obtained are applied to various situations in order to estimate the isotopic changes during water flow through a basin.

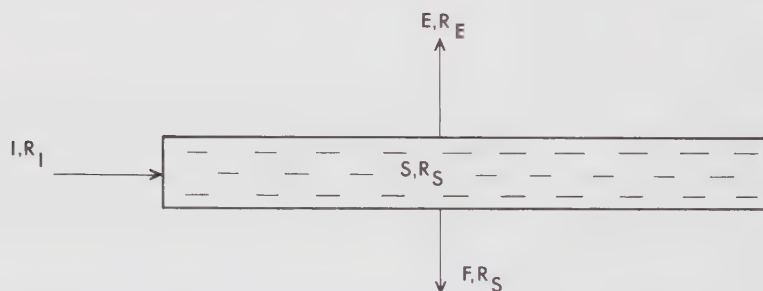


Fig.6.1 Isotope balance for a well-mixed water body. I, E, F, and S are the inflow, evaporation, outflow, and mass of water respectively, and the R symbols the corresponding isotope ratios.

$$\text{Mass balance} \quad \frac{dS}{dt} + E + F - I = 0 \quad (6.8)$$

$$\text{Isotope balance} \quad \frac{d(S R_S)}{dt} + E R_E + F R_S - I R_I = 0 \quad (6.9)$$

where S = mass of water

t = time

E = evaporation rate

F = outflow, for example infiltration

I = inflow

R_S , R_E and R_I = the isotopic ratios of the water body, evaporate and inflow respectively

Integration of Eq.(6.8) yields

$$S = S_0 - \int_0^t (E + F - I)dt \quad (\text{for } \int_0^t (E + F - I)dt < S_0) \quad (6.10)$$

where S_0 = mass of water at $t = 0$

Eqs.(6.8), (6.9), and (6.10) give

$$\frac{dR_S}{dt} = \frac{E(R_S - R_E) - I(R_S - R_I)}{S_0 - \int_0^t (E + F - I)dt} \quad (6.11)$$

or, in the δ -notation

$$\frac{d\delta_S}{dt} = \frac{E(\delta_S - \delta_E) - I(\delta_S - \delta_I)}{S_0 - \int_0^t (E + F - I)dt} \quad (6.12)$$

If the various water flows and the isotopic composition of the inflow are known, Eq.(6.12) can be solved numerically for δ_S , using δ_E from Eq.(6.5). For δ_E , values of δ_a and h ($= e_a/e_s$) are needed as well as the isotopic fractionation factors.

Some special cases can be considered:

$$1. \text{ No inflow. Then } \frac{d\delta_S}{dt} = \frac{E(\delta_S - \delta_E)}{S_0 - \int_0^t (E + F)dt} \quad (6.13)$$

Enrichment will proceed until $\delta_S = \delta_E$.

2. No outflow, evaporation equals inflow (constant volume)

$$\frac{d\delta_S}{dt} = \frac{I}{S_0} (\delta_I - \delta_E) \quad (6.14)$$

Enrichment will proceed until δ_S reaches a value such that $\delta_E = \delta_I$.

Eq.(6.12) is easy to handle and the physical meanings of the different terms are readily seen, as exemplified by the special

cases above. However, the use of δ_E is possible only when $E \neq 0$. When $E = 0$ then $h = 1$ and δ_E is not defined. Because of mass exchange there may well be a net isotopic flux through the surface also when the net evaporation is zero.

This problem can be overcome by expressing the isotope flux by net evaporation, $E R_E$, in the isotope balance Eq.(6.9) directly as E_i by Eq.(6.1). Using the evaporation rate from Eq.(6.2), the expression for isotopic change of a water body Eq.(6.12) then becomes

$$\frac{dR_s}{dt} = \frac{\frac{C}{r} [e_s R_s (1 - \frac{1}{\alpha\beta}) - e_a (R_s - \frac{1}{\beta} R_a)] - I(R_s - R_E)}{S_0 - \int_0^t (\frac{C}{r} (e_s - e_a) + F - I) dt} \quad (6.15)$$

or, in the δ -notation

$$\frac{d\delta_s}{dt} = \frac{\frac{C}{r} [e_s(\delta_s+1)(1 - \frac{1}{\alpha\beta}) - e_a(\delta_s+1 - \frac{1}{\beta}(1+\delta_a))] - I(\delta_s - \delta_I)}{S_0 - \int_0^t (\frac{C}{r} (e_s - e_a) + F - I) dt} \quad (6.16)$$

As in Eq.(6.5), e_a in Eq.(6.16) can be replaced by $h \cdot e_s$.

It is to be noted that the kinetic fractionation, as reflected by β , is present in Eq.(6.16) also during condensation. This is contrary to common assumptions. No reason could be found to disregard kinetic fractionation in the atmospheric laminar sub-layer when the net flow is directed into the liquid.

Two slightly different equations by which the isotopic change of a water body can be calculated have now been derived. Eq.(6.12), with δ_E from (6.5), is suitable when E is known explicitly, for instance from water-budget considerations, or when E can be given a reasonable value from "general experience". One advantage with Eqs.(6.12) and (6.5) is that δ_E can be roughly estimated even if the input variables for Eq.(6.5) are not measured (cf. Section 6.1.3). In more general considerations, when the net evaporation may be positive, zero, or negative, Eq.(6.16) is preferable. This is the case when isotopic changes of a snowpack are studied, for instance.

Simulations with Eqs.(6.12) and (6.5) or (6.16) are made in Section 6.2 in order to examine enrichment rates to be expected within a basin.

The rate of turbulent mixing in the atmosphere, as expressed by the wind function, C/r , might be known for lakes, either determined empirically using evaporation rates determined by other methods or derived from the wind profile. For water bodies on the forest floor or in clearings, as in the present study, C/r is not well known since the logarithmic wind profile is not valid, and it is difficult even to get accurate values of the wind speed. Therefore, when the wind function is needed in simulations, its value is determined from assessed values of E , e_s , and $1 - h$, under conditions of positive net evaporation.

6.1.3 Sensitivity of the fractionation model to meteorological conditions

Before the fractionation model is applied to situations related to the present runoff study, the process of isotopic change of a water body under different meteorological situations will be discussed. Simulations with Eqs.(6.5) and (6.12) are performed for an initially 100-mm-deep water body without inflow or outflow, using constant values for the meteorological variables. The situation is intended to represent Swedish summer conditions. Reference values and realistic ranges of the variables are given in Table 6.1, which also shows estimated ranges of the enrichment to be discussed below. The water temperature is assumed equal to the air temperature, making h in Eq.(6.5) equal to the relative humidity of the air.

Table 6.1 Reference values, ranges of variables and estimated enrichment variables for a 100 mm deep water body.

Variable	Reference value	Range	Steady state value ¹ (‰)	Rate of enrichment (‰ day ⁻¹ during the first day)
Reference situation			2.5	0.82
Evaporation (E)	3 mm day ⁻¹	1-5 mm day ⁻¹	2.5	0.27-1.35
Relative humidity (h)	70%	50-90 %	18.4 - -6.2	0.80-0.88
$\delta^{18}O$ of the atmosphere (δ_a)	-20‰	-25 - -15‰	-2.7-7.6	0.48-1.15
Initial $\delta^{18}O$ of the water	-10‰	-15 - -5‰	2.5	0.48-1.15
α	1.011 (+10°C)	1.010-1.012 (+20°C) (0°C)	1.0-3.9	0.72-0.90
β	1.016	1.014-1.018	1.6-3.3	0.76-0.89

¹ Defined in the text below

As pointed out earlier, the $\delta^{18}\text{O}$ of the atmospheric water vapour, δ_a , is needed when calculating the isotopic changes of water exposed to the atmosphere. Measurements of δ_a in Sweden have been reported by Saxena (1987). From 4 years of continuous measurements he showed that rainwater (in contrast to falling snow) generally is close to isotopic equilibrium with the (saturated) atmosphere during the rainfall event. At such equilibrium we have by Eq.(3.11)

$$R_p = \alpha R_a$$

giving

$$\delta_p = \alpha(\delta_a + 1) - 1 \approx \delta_a + \alpha - 1$$

or

$$\delta_p - \delta_a \approx \alpha - 1$$

At 10°C $\alpha = 1.011$ giving $\delta_p - \delta_a \approx 11\text{‰}$. The water body in Table 6.1 is assumed to be newly fallen rainwater, with $\delta^{18}\text{O} = -10\text{‰}$. The reference value of δ_a , -20‰ , is chosen so as to roughly represent equilibrium during the rainfall.

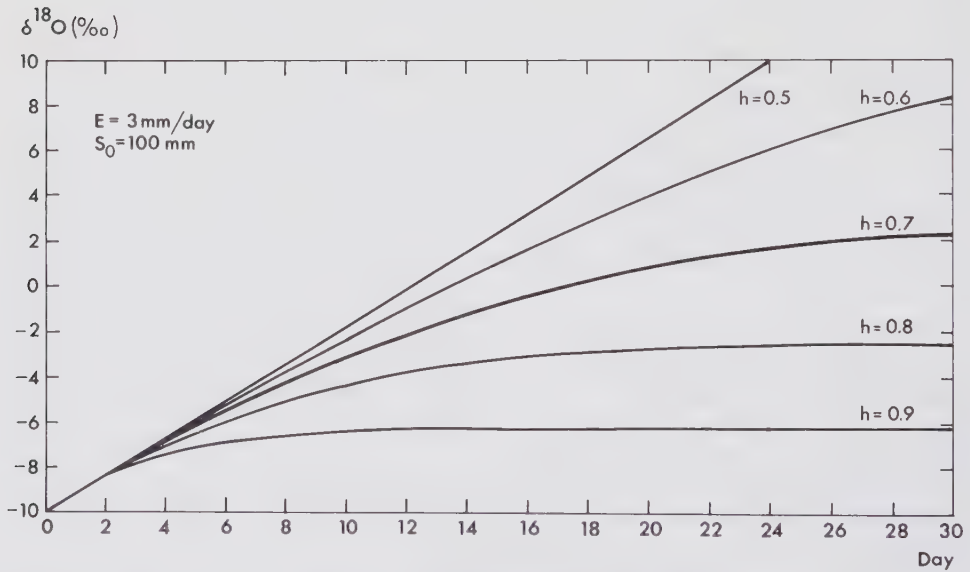


Fig.6.2 Simulation of the $\delta^{18}\text{O}$ of water under reference enrichment conditions (the relative humidity $h = 0.7$) and sensitivity to variations in the relative humidity.

Fig.6.2 shows calculated isotopic change in the water under reference conditions ($h = 0.7$) and also for various values of h . Enrichment proceeds with a decreasing rate as δ_s approaches a steady-state value where $\delta_s = \delta_E$ (Eq.6.13). For $h = 0.7$ the steady-state value is reached when $\delta_s - \delta_a = 22.5\text{‰}$, which occurs after about one month. At that time only 10 mm remains, since 90 mm has evaporated.

The enrichment of a water body can be characterized by the steady-state value and by the rate of change. Table 6.1 shows how the ranges of each individual variable affect the steady-state value and the rate of enrichment expressed as enrichment during the first day, with the other variables having their reference values. The sensitivity is discussed in more detail below in an attempt to clarify the process of isotope fractionation in water.

The steady-state value

The steady-state value is defined here as the isotopic equilibrium value at the prevailing relative humidity and δ -value of the atmosphere. The term "steady-state value" is preferred to "equilibrium value" in order to reserve "equilibrium value" for the case of thermodynamic equilibrium between water and vapour (100% relative humidity).

From Eq.(6.13) it is seen that isotopic change in a water body without inflow ceases when $\delta_s = \delta_E$. Eq.(6.5) then gives for the steady-state value, $\delta_{s,s}$,

$$\delta_{s,s} - \delta_a = (1 + \delta_a) \frac{\frac{1}{\alpha} - h - \beta(1 - h)}{\beta(1 - h) - \frac{1}{\alpha}} \quad (6.17)$$

This relationship, which gives the maximum possible enrichment at a certain δ_a as a function of the relative humidity, is plotted in Fig.6.3.¹ The diagram also shows the conditions for enrichment or depletion during evaporation. For combinations of $\delta_s - \delta_a$ and h that fall below the curve, as in the example discussed below, enrichment will take place. Points above the curve characterize situations where evaporation causes depletion. The

¹ As is seen in Eq.(6.17), $\delta_{s,s} - \delta_a$ is a function also of δ_a , but realistic ranges of δ_a result in no visible deviation from the curves in Fig.6.3.

latter situations are probably rare for longer time periods in natural waters, but may occur at night and during late autumn (low δ_a , high h). Conditions for enrichment are discussed in Section 6.2.6 when $\delta^{18}\text{O}$ of snow is analysed. Here, only enrichment situations are considered.

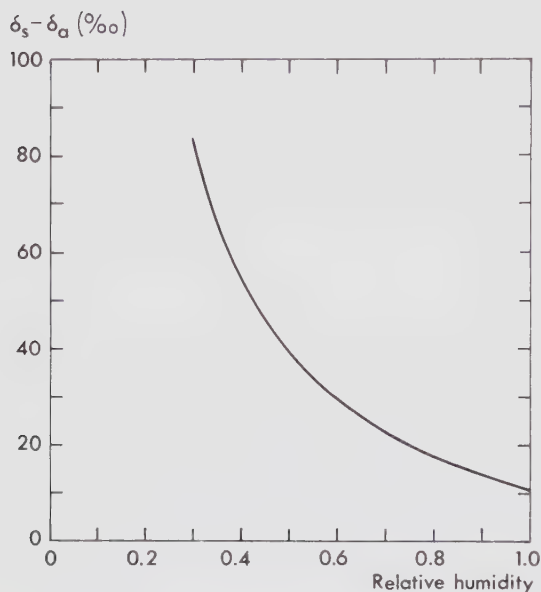


Fig.6.3 Difference between the $\delta^{18}\text{O}$ of water (δ_s) and atmosphere (δ_a) at "steady-state" conditions as a function of the relative humidity.

At 100% relative humidity no evaporation occurs, but molecular exchange proceeds until the water and atmosphere are in isotopic equilibrium, i.e., $\delta_s - \delta_a \approx 11\text{‰}$. With decreasing relative humidity the maximum enrichment for a certain δ_a increases, since the ^{18}O flow from the atmosphere gradually decreases due to decreasing concentration of H_2^{18}O in the atmosphere.

The great sensitivity of the steady-state value to variations in h , particularly at low values of h , is evident from Fig.6.3. A change of one percentage point in h changes $\delta_{s,s}$ by 0.30‰ at $h = 95\%$ and by 0.95‰ at $h = 55\%$.

The difference between the $\delta^{18}\text{O}$ of water and of the evaporating vapour

From Eq.(6.12) it is seen that the rate of enrichment is directly proportional to E and to $\delta_s - \delta_E$ and inversely proportional to

$$S = S_0 - \int_0^t E \, dt.$$

When examining the influence of $\delta_s - \delta_E$ on the rate of enrichment, the factors determining $\delta_s - \delta_E$ must be considered. From Eq.(6.5) the following expression for the difference between the δ -value of the water surface and that of the net evaporate is obtained

$$\delta_s - \delta_E = \frac{\delta_s[\beta(1-h) - \frac{1}{\alpha}] - \frac{1}{\alpha} + h(\delta_a + 1) + \beta(1-h)}{\beta(1-h)} \quad (6.18)$$

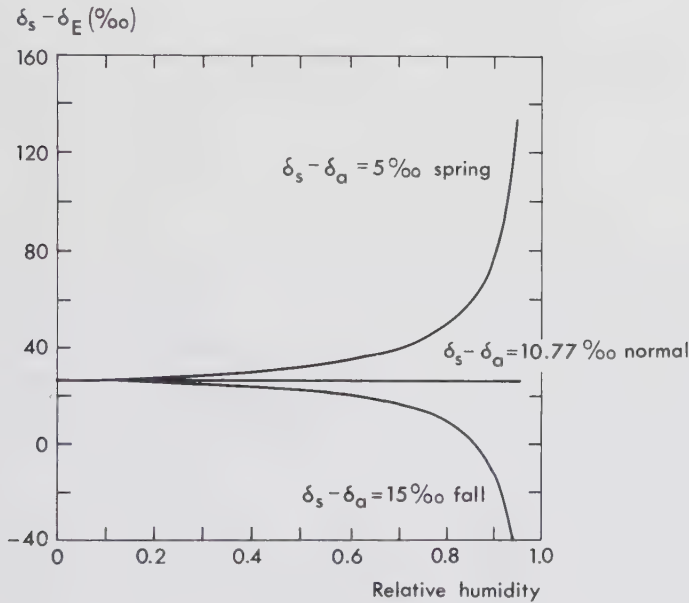


Fig.6.4 Difference between the $\delta^{18}\text{O}$ of water (δ_s) and net evaporate (δ_E) as a function of the relative humidity. The difference is given for various differences between the $\delta^{18}\text{O}$ of water and atmosphere (δ_a) to be expected for surface water in a basin.

This difference is plotted against h for selected values of $\delta_s - \delta_a$ in Fig.6.4¹. It can be concluded from the figure that for a certain δ_s , the difference $\delta_s - \delta_E$ (and thus the rate of enrichment at a given evaporation rate) is very sensitive to δ_a and h when h is high. At low h (<60%) the difference depends

¹ As seen in Eq.(6.18), the difference is also a function of δ_s , but realistic ranges of δ_s cause very small changes in the curves.

very little upon δ_a and h . As h approaches zero, $\delta_s - \delta_E$ approaches $\alpha\beta - 1$ ($= 27\text{‰}$). This is obtained directly from Eq.(6.3) giving $R_E = R_S/\alpha\beta$ for $h = 0$.

It should be noted that the above comments concern the dependence of the rate of enrichment on h for given values of E . But for a given rate of turbulent exchange and surface temperature, E decreases as h increases. The large values of $\delta_s - \delta_E$ thus occur in connection with small values of E . A closer analysis shows that for constant turbulent exchange and surface temperature, the rate of enrichment decreases linearly with increasing values of h . The larger the difference $\delta_s - \delta_a$, the larger is the decrease with h .

Fig.6.4 can be used to obtain values of $\delta_s - \delta_E$ in rough estimates of the rate of enrichment by Eq.(6.12). It is seen that the dependence of $\delta_s - \delta_E$ upon h is a function of $\delta_s - \delta_a$. When $R_a = R_s$ $1/\alpha$, i.e., $\delta_s - \delta_a = 11\text{‰}$, we get by Eq.(6.3) $R_E = R_S/\alpha\beta$, or $\delta_s - \delta_E = 27\text{‰}$ for all values of h . This is the initial value of $\delta_s - \delta_E$ to be expected for a rainwater puddle starting to evaporate after the rainfall when h drops but δ_a remains constant for some time. As δ_s increases by the enrichment, $\delta_s - \delta_a$ increases, making $\delta_s - \delta_E$ decrease until the steady state is reached, when $\delta_s - \delta_E = 0$. The points of intersection of the curves with $\delta_s - \delta_E = 0$ correspond to the curve for the steady-state value in Fig.6.3. From the seasonal variation of δ_a , which largely follows that of air temperature, it can be expected that $\delta_s - \delta_a < 11\text{‰}$ for instance for meltwater in spring and streamwater in summer. In such cases, the rate of enrichment for a certain evaporation rate is comparatively high, since $\delta_s - \delta_E$ is large. Values of $\delta_s - \delta_a > 11\text{‰}$, giving a comparatively low rate of enrichment, may occur for streamwater in late autumn.

6.1.4 Experimental results related to the present study

Saxena (1987) found that the isotopic changes of water in 0.11 m^2 evaporation pans were well described by Eqs.(6.12) and (6.5). The experiment was performed in Uppsala, Sweden, which has a climate representative of the present runoff study. Pan water was sampled with time intervals of about 1-3 days. As an example is shown measured and calculated $\delta^{18}\text{O}$ of a pan without inflow, having an initial depth of 95 mm (Fig.6.5). The calculations by Eqs.(6.12) and (6.5) were performed using period

mean values of h (assuming surface temperature = air temperature) and E (obtained from water budget). Under real conditions, the steady-state value is never reached since it continuously changes with h and δ_a . In the later part of the experiment the δ -value of the pan water approaches, but never reaches, the meteorologically determined steady-state values.

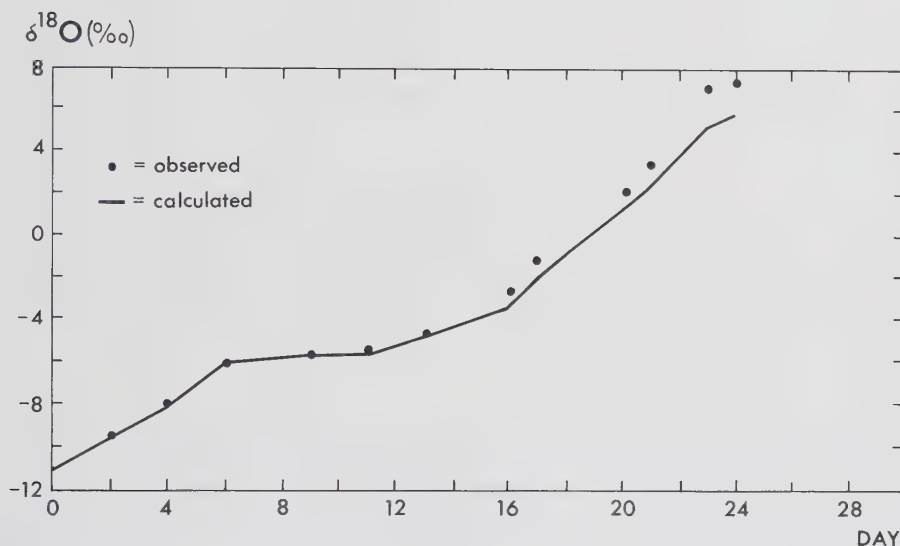


Fig.6.5 Observed and calculated $\delta^{18}\text{O}$ of water in an evaporation pan having an original water depth of 95 mm. Uppsala, June-July 1982. (From Saxena, 1987).

6.1.5 Limitations of the model for isotopic change

In the model for isotopic change in a water body, complete mixing of the water was assumed. The net isotopic flux through the surface was assumed to change the isotopic composition of the whole mass of water at the same rate. This assumption is justified for stream channels and probably also for shallow water reservoirs, at least when considering periods of several days, since complete mixing regularly occurs due to nocturnal cooling. There are, however, many situations within a basin in which the effects of fractionation on the bulk mass of water are reduced by incomplete mixing.

An instructive example is evaporation from ice, a process that is non-fractionating. Here the transport of molecules to the surface occurs only by molecular diffusion, and it is too slow

to allow the lighter molecules to reach the surface at a higher rate than the heavier ones. The evaporation takes away the molecules present in the surface layer, as a cheese cutter slices off the cheese. Enrichment to a steady-state value¹ will rapidly occur at the surface, but it will not affect the composition of the bulk ice. Stated in another way: the molecular diffusion is too slow to allow the enrichment to proceed inwards. The situation can be described by Eq.(6.14). A very small reservoir, S_0 , is moving into the ice at a rate equal to the evaporation. The reservoir "sees" an inflow, I , equal to E , having the isotopic composition of the ice, δ_s . By Eq.(6.14) we get $\delta_E = \delta_I = \delta_s$.

Other phase transitions where fractionation is prevented or diminished due to incomplete mixing are transpiration, evaporation from soil, evaporation from snow, and also melting of snow. These cases are discussed in the next section, together with cases of complete mixing, in attempts to map the possibilities of enrichment during water flow through a basin.

6.2 Estimates of enrichment rates during water flow through a basin

This section deals with fractionation of water on its way from precipitation to the stream and in the stream itself. Except for the studies on snowpacks referred to below, there are only few studies reported in the literature on fractionation during water flow through a basin.

Enrichment of rainwater during groundwater recharge (interception and infiltration) in an arid area were discussed by Gat and Tzur (1967) using a simplified version of the Craig and Gordon model (1965), which allowed analytical solutions of the total enrichment. There are also a few studies on fractionation in the unsaturated zone, some of which will be referred to below. No reports treating fractionation during overland flow or streamflow have been found. As is shown in this section, situations may occur in which there is a significant isotopic change during these latter processes, a change that should be taken into account when interpreting the isotope content of streamwater.

¹ This steady-state value is different from that given by Eq.(6.17). Apart from h and δ_a it also depends on the rate of evaporation and the composition of the ice.

At the present stage it is not considered possible to estimate the combined result of the various fractionations taking place during water flow through a basin. The aim of the following paragraphs is rather to identify fractionation situations and to suggest possible ways of estimating enrichment or depletion rates to be anticipated at various locations along the path of the water flow. Implications for hydrograph separation by ^{18}O are discussed in the uncertainty discussion in Chapter 9.

This section is arranged following the flow of water through a basin, discussing fractionation during interception, infiltration, soil water flow, groundwater flow and flow in discharge areas including the stream channel. The flow pattern is broken by the discussion on snow, which comes last in the section. When ^{18}O is used in modelling the water flow through a basin, possible enrichment at any stage is important, whereas for hydrograph separation possible enrichment at certain stages is of less importance than at others.

For some stages of the water flow such as soil water, groundwater and snow, the discussion is based mainly on results from the literature. For water on the ground surface or in the stream, possible rates of enrichment are estimated using field data from the present investigation as examples. These examples are chosen to give the upper limits of enrichment to be expected.

In the attempts to quantify enrichment rates in connection with rainfall or snowmelt, the following assumptions are made, if not otherwise stated:

The falling raindrops are in isotopic equilibrium with the saturated atmosphere, i.e., $\delta_p = \alpha(1 + \delta_a) - 1$. For $\delta_a = -20\text{‰}$ and $\alpha = 1.011$ this gives $\delta_p = -9.22\text{‰}$. Due to heat stored in the vegetation or in the ground, the atmosphere may not be completely saturated with respect to the wet surfaces. This allows some evaporation and thereby enrichment during the rainfall. At $+10^\circ\text{C}$, a temperature difference of 0.3°C between surface and air gives a vapour-pressure ratio of 98%. The enriching effect of this is examined in some of the examples. After the cessation of rainfall, the same δ_a is assumed as during the rain, and the relative humidity is assumed to have dropped to 70% or, in the case of interception, to 80%. During snowmelt, isotopic equilibrium between meltwater and the atmosphere cannot be expected to prevail. The seasonal variation in δ_p suggests values

of δ_a during snowmelt around -20‰ , and $\delta^{18}\text{O}$ of meltwater, δ_M , around -15‰ . A relative humidity of 70% is assumed.

Values of meteorological and other input variables needed in the enrichment models are often taken from "general experience" in an attempt to avoid overburdening the text with references and detailed discussions on each assumption.

The magnitudes given below of the estimated isotopic changes in different situations should be related to the yearly variations in the δ -values found within a basin. In Fig.3.3 it was shown that the precipitation $\delta^{18}\text{O}$ may vary from -25 to 0‰ over the year. Experience from the present study shows that at these precipitation δ -values, streamwater and groundwater $\delta^{18}\text{O}$ may vary between about -13 and -9‰ . Let us also recall that the $\delta^{18}\text{O}$ of a water sample is determined with an accuracy of about $\pm 0.2\text{‰}$.

6.2.1 Interception

From theoretical considerations treating an arid area, Gat and Tzur (1967) concluded that the mean ^{18}O enrichment of the throughfall never exceeded 0.5‰ .

The only reported measurements of enrichment by interception are those by Saxena (1986, 1987), performed in a dense pine forest in Uppsala. During a four-month summer period he found that about 40% of the rainfall was lost by interception. The weighted mean $\delta^{18}\text{O}$ of the throughfall was enriched by 0.3‰ as compared to the precipitation. Saxena (1986, 1987) applied the present enrichment model to the case of interception and found that the estimated enrichment largely agreed with observed values.

Saxena observed both enrichment and depletion of the throughfall during single rainfall events. The directions of isotopic change could be related to the observed difference between the δ -values of the rainfalls and the atmosphere. The isotopic change in the 24 reported storms ranged from -0.4 to 1.2‰ , with a median value of 0.2‰ .

The total interception loss at this experiment was considerably larger than what has been reported from other studies in Swedish coniferous forests, 20-25% of summer rainfall (Bringfeldt and Hårsmar, 1974; Perttu et al., 1980). It seems probable

therefore that the isotopic changes are generally smaller than those reported by Saxena.

6.2.2 Infiltration

Non-ponding infiltration

Isotopic change in infiltrating water is estimated by applying Eqs.(6.12) and (6.5) to a very shallow, hypothetical, reservoir fed by rainwater or meltwater, and drained by infiltration (and evaporation). The reservoir depth, S_0 , affects only the rate of isotopic change and not the maximum enrichment. By choosing a sufficiently small S_0 , say 0.1 mm, stationary conditions are rapidly achieved and the calculated enrichment is independent of S_0 .

The assumption of isotopic equilibrium between rainwater and a saturated atmosphere implies no enrichment during rainfall infiltration. If, on the other hand, the normalized relative humidity is slightly below 100%, a non-negligible enrichment may occur at low rainfall intensities. Calculated enrichment for $h = 98\%$ is plotted against P/E (P = precipitation rate) in Fig.6.6 (unbroken line). Due to low rates of turbulent mixing close to the the forest floor the evaporation rate is probably smaller than 0.1 mm h^{-1} , which, for rainfall intensities larger than 1 mm h^{-1} , gives $P/E > 10$ and a maximum enrichment of about 0.5‰ .

For infiltration during snowmelt the possibilities for enrichment are somewhat different, since the conditions are far from isotopic equilibrium. The broken line in Fig.6.6 shows calculated enrichment at meltwater $\delta^{18}\text{O} = -15\text{‰}$, $\delta_a = -20\text{‰}$ and $h = 70\%$. (P/E in the figure is replaced by M/E , where M = rate of meltwater flow from the snowpack). The enrichment for a certain P/E is considerably larger in this case, but E is very small, since the hypothetical reservoir is covered by snow. The value of M seldom exceeds a few mm h^{-1} , but since the magnitude of E is not known, quantitative conclusions on the enrichment cannot be drawn. (For example $E = 0.01 \text{ mm h}^{-1}$ and $M = 1 \text{ mm h}^{-1}$ gives $\Delta\delta \approx 0.5\text{‰}$.)

In connection with the present investigation, a field experiment was performed showing no enrichment during the flow of rainwater on or within the top layer of the soil.

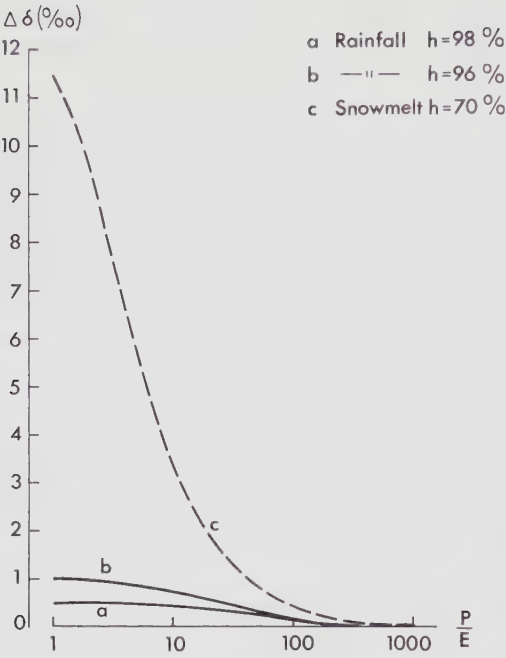


Fig.6.6 Calculated enrichment ($\Delta\delta$) during non-ponding infiltration as a function of the ratio between the rates of precipitation (P) (rainfall or snowmelt) and evaporation (E).

The experiment was performed in Lagga, ca 15 km east of the Nåsten basin, with 1 m wide by 5 cm deep horizontal troughs inserted in vertical pit walls, about 5 cm below ground surface, in or just below the humic layers. $\delta^{18}\text{O}$ and electrical conductivity of water collected in the troughs during a rainfall are shown in Table 6.2.

Table 6.2 $\delta^{18}\text{O}$ and electric conductivity of water collected in surface-near troughs.

Trough No.	26/8 1979 14.00 h			26/8 1979 19.30 h		
	$\delta^{18}\text{O}$ (‰)	γ mS m ⁻¹	q'_s ¹ (mm)	$\delta^{18}\text{O}$ (‰)	γ mS m ⁻¹	q'_s (mm)
1	-9.6	4.7	1	-8.4	3.3	1
2	-9.5	4.5	1	-8.4	2.9	1
3	-9.5	7.5	1	-8.9	5.0	1
4	-9.3	6.5	0.1	-8.4	3.6	0.1
5	-9.7	23	1	-8.7	14	1
Rainfall	-9.7	2.5	29 mm	-8.5	2.1	23 mm

¹ Collected amount per unit of catchment area.

Troughs Nos. 1 - 4 were in forested till soil, whereas No. 5 was in heavy clay at a grassy, formerly agricultural site. The collected water volumes constituted only small fractions of this prolonged rainfall of medium intensity (4 mm h^{-1}). The collected water probably infiltrated just above the collectors and was caught on its downward-downslope flow in the humic layers. It may thus represent infiltrated water rather than overland-flowing water. The $\delta^{18}\text{O}$ showed no significant difference from that of rainfall. The conductivity of the water, on the other hand, increased considerably, particularly at the clayey site.

Infiltration from surface reservoirs¹

In forested till soils in Sweden, surface-water reservoirs are mainly restricted to discharge areas, since the infiltration capacity of the soil is generally greater than the intensity of rainfall or snowmelt. Enrichment in surface reservoirs in discharge areas does not affect the isotopic composition of groundwater. This subject is treated later, in connection with enrichment in stream channels. Here the enrichment during the less probable case of water infiltration from surface reservoirs generated by rainfall or snowmelt excess is estimated.

It is assumed that isotopic equilibrium between the rainwater and the saturated atmosphere prevails during the rainfall, implying no enrichment in the reservoir. After the rainfall, the relative humidity is assumed to have dropped to 70%.

Mean enrichment of the infiltrated water, calculated by Eqs. (6.12) and (6.5), is plotted against the infiltration rate for different evaporation rates in Fig.6.7. As was pointed out by Gat and Tzur (1967), the mean enrichment is independent of the initial water depth, since the longer time for evaporation and molecular exchange of a deep reservoir is exactly compensated for by the larger mass of water in which the enriched surface water is mixed. The curves in Fig.6.7 agree well with analogous theoretical curves presented by Gat and Tzur.

Infiltration capacities in till soils are probably seldom lower than 20 mm h^{-1} and commonly considerably higher. One could therefore hardly expect an enrichment of more than a few tenths

¹ Throughout this text the term "surface reservoir" is used for any collection of water on the ground: pools, ponds, water in flooded areas etc.

of per mil during rainfall ponding on an unsaturated soil. The ponding requires very high rainfall intensity, accounting for a small fraction of total rainfall and infiltration. The enrichment effect of rainfall ponding on the isotopic composition of soilwater and groundwater is therefore probably negligible.

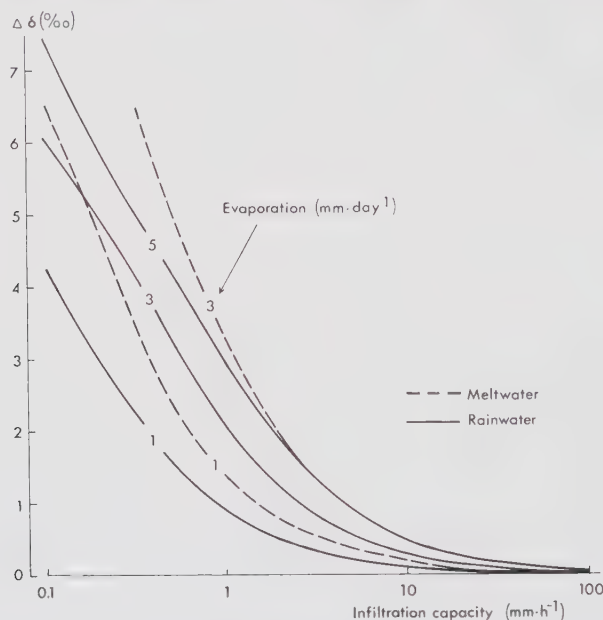


Fig.6.7 Calculated mean enrichment ($\Delta\delta$) of water infiltrating from a surface-water reservoir as function of the infiltration capacity of the soil.

In calculating the mean enrichment of the infiltrating meltwater in Fig.6.7, the same enrichment variables as those for meltwater in Fig.6.6 have been used (δ_s originally -15‰ , $\delta_a = -20\text{‰}$, $h = 70\%$). As in Fig.6.6, the calculated enrichment for meltwater is considerably larger than for rainwater. However, the larger mean enrichment for a certain evaporation rate is compensated for by a lower evaporation rate during snowmelt. Evaporation is perhaps 1 mm day^{-1} for water reservoirs exposed to the atmosphere and much smaller for water standing in the snowpack. The effect of enrichment is therefore most likely negligible in this case as well.

Enrichment in surface-water reservoirs in discharge areas not connected to the channel network may affect the isotopic composition of groundwater under certain circumstances. In slightly undulating terrain of low mean slope, with very shallow groundwater, the groundwater table may intersect the ground surface

at several localities. This results in the existence of many surface reservoirs without surface outlets. In one of the present study areas, the Nåsten basin, groundwater pools are frequent at high groundwater stages. The pools are emptied by evaporation and mainly lateral drainage, the latter causing infiltration of enriched surface water. The fall in water level in such pools during late spring is, say, 1 cm day^{-1} (0.4 mm h^{-1}), of which evaporation accounts for $1\text{--}2 \text{ mm day}^{-1}$. In Fig.6.7 it is seen that the mean enrichment may reach several per mil. But since this infiltration occurs only locally, its effect upon groundwater $\delta^{18}\text{O}$ in a basin is not likely to be very important.

6.2.3 Soil water

One of the main methodological problems in performing isotopic hydrograph separations is to get a good estimate of the ^{18}O content of the groundwater component of streamflow. As is discussed in Section 8.1.4, the groundwater discharging to a stream may partly consist of water that was stored in the unsaturated zone close to the stream before the event. Due to the seasonal variation in the $\delta^{18}\text{O}$ of precipitation, the ^{18}O content of soil water often differs from that of groundwater. A possible enrichment in the unsaturated zone may, however, introduce systematical errors in the assessment of groundwater $\delta^{18}\text{O}$ during runoff events.

The isotopic composition of soil water may be modified by direct evaporation and exchange with the atmosphere or by transpiration. In both cases, the isotopic changes of the bulk volume of soil water are greatly reduced by incomplete mixing with the enriched surface layer (cf. Section 6.1.5).

The effects of evaporation from soil and transpiration are treated separately below, after which some experimental results found in the literature are presented.

Bare soil

Direct evaporation from soil water is of significance mainly in bare soils. Since the soil air within a wet soil is very close to saturation, evaporation takes place only from the uppermost wet layer, which is supplied by water flow from below. The evaporating layer strives to reach the isotopic "steady-state value" with respect to the atmosphere.

As is pointed out by Dinger et al. (1974), the enrichment in the top layers may be transported downwards by molecular diffusion through liquid and vapour. In addition to these molecular flows, the whole enriched water layer is pushed down during periods of infiltration and downward water flow. The effect of this latter, advective, transport depends on the frequency of infiltration situations in relation to the rate of enrichment of the surface layer.

Zimmermann et al. (1967) treated, experimentally and theoretically, the enrichment penetration in a saturated soil evaporating at a constant rate at the surface. They showed that the enrichment distribution with depth is exponential, with the mean depth of enrichment, z , and the adjustment time, τ , given by

$$z = \frac{D}{v} \quad (6.19)$$

$$\tau = \frac{D}{v^2} \quad (6.20)$$

where D = molecular diffusion coefficient for $H_2^{18}O$ in water.

At $+10^\circ C$, $D = 1.7 \cdot 10^{-9} \text{ m}^2\text{s}^{-1} = 1.5 \text{ cm}^2 \text{ day}^{-1}$
(Wang 1965).

v = upward velocity of water $\approx E/p$, where p = porosity
and E = evaporation rate.

It is seen from the equations that the greater the evaporation rate, the smaller is the penetration depth and the shorter the time to reach steady state. For evaporation rates of 1 and 3 mm day $^{-1}$ Zimmermann et al. calculated mean penetration depths of 0.15 and 0.05 m, respectively, and adjustment times of 150 and 17 days, respectively.

The case of enrichment penetration in unsaturated soil is more complicated, due to the various flow mechanisms involved. Münnich et al. (1980) presented a model to explain the very low slope of ^{18}O -D line found by Dinger et al. (1974) in drying Saharan sand dunes. Allison (1982) concluded, from laboratory experiments, that low slope is caused by an extra kinetic separation taking place in drying sand during the molecular diffusion of water vapour through a dry top layer.

Transpiration

Transpiration is probably non-fractionating (Zimmermann et al. 1967, Allison et al. 1984), but few experiments have been reported on this question. Conceivable points where fractionation could occur during water flow through vegetation are at the root-soil water boundary and at the transpiring leaves.

No evidence of selective isotopic water uptake has been found in the literature. Zimmermann et al. reported no isotopic changes of the remaining water during transpiration from a vase, i.e., from a very large saturated pore. Fractionation in such a case is not very probable, since there are no phase transitions present. Selective isotopic water uptake during unsaturated conditions, due to some diffusive or biological process, cannot, however, be excluded based on the reported experimental data. Allison et al. (1984) found no, or possibly a very small, enrichment of soil water by transpiration in a drying soil. They pointed out the possibility for fractionation during water uptake by roots in very dry soils, since vapour transport then may be important.

As in the case of evaporation from soil, considerable enrichment probably takes place at the evaporating surfaces of leaves. Gonfiantini et al. (1965) found an ^{18}O enrichment of about 10‰ in transpiring leaves as compared to the soil water. The molecular diffusion of H_2^{18}O in liquid H_2^{16}O is, however, too slow to allow the enrichment to proceed against the water flow in the stems. Upward water velocity through the stems of a transpiring forest is on the order of $0.1\text{--}10\text{ m h}^{-1}$ (Kramer and Kozlowsky 1979). The corresponding mean penetration depths from Eq.(6.19) with $D = 1.7 \cdot 10^{-9}\text{ m}^2\text{ s}^{-1}$ are fractions of millimetres. Since the estimated flow velocities are for stems only, the estimates do not exclude deeper penetration into the leaves, but it is evident that no enrichment can penetrate against the water flow from the canopy to the soil water. Downward enrichment flow by advection, important in bare soils, is moreover not possible during transpiration, since the water flow in the stem is never directed downwards.

It can be concluded that the only stage of the water flow through vegetation which could have fractionating effect on the soil water is the root-soil water boundary. No evidence of fractionation at this boundary has been found in the literature, and since it would occur only when vapour flow to the

roots becomes important, it would be of little significance in humid areas.

Experimental results reported in the literature concerning enrichment of soil water

The mean isotope content of the water in the soil profile and that of the percolate may well differ from that of water input, even if no fractionation takes place in the soil. The transpiration removes some of the infiltrated water from the soil. What part of the water it takes depends on the vertical distribution of soil water tension and root density, but not on the isotope content of the water. If the mean $\delta^{18}\text{O}$ of the transpired water differs from the mean $\delta^{18}\text{O}$ of the water input, the mean $\delta^{18}\text{O}$ of the percolate will differ from that of the water input. It may thus be difficult to draw conclusions on the rate of enrichment in soil water from comparisons between the $\delta^{18}\text{O}$ of soil water and percolate and the $\delta^{18}\text{O}$ of precipitation. Possible enrichments during snow storage, interception, and infiltration also have to be regarded in such comparisons.

In view of the extensive use of stable isotopes in groundwater studies, surprisingly few investigations of stable isotopes in the unsaturated zone have been reported. The reported studies include soil-water sampling, lysimeter experiments, and laboratory experiments.

The seasonal variations of precipitation ^{18}O was found in the vertical distribution of soil water ^{18}O by Saxena (1984). He investigated soil profiles in glaciofluvial sand (Uppsala) and till soil (close to the Buskbäcken basin); both sites being forested. The observed variations of soil water ^{18}O correspond to a damping of the variations in the monthly precipitation values of about 20-40 %. Identification of ^{18}O depleted melt-water at two depths made it possible to estimate the rate of percolation during the in-between year. The weighted mean $\delta^{18}\text{O}$ of one year's soil water below the root zone differed very little from that of precipitation during the same period. No conclusions concerning the extent of enrichment were drawn.

Seasonal variations of precipitation δ -values were also found in deuterium profiles of soil water in Western Germany by Eichler (1965). Certain profiles showed heavy enrichment in the top 10 cm as compared to the precipitation.

Zimmermann et al. (1967) compared deuterium content of soil water below bare and grass-covered surfaces in Western Germany. Assuming that transpiration is non-fractionating, they found a mean enrichment below bare soil of $\Delta\delta D$ of about 10‰, corresponding to $\Delta\delta^{18}O$ about 2‰.¹ The higher δ -values below the bare soil were found in the whole sampled depth interval, 0-2 m.

Low slopes of D- ^{18}O -plots of soil water in sand dunes indicating enrichment by evaporation were reported by Dinger et al. (1974), but no figures on the magnitude of the enrichment of the infiltrated water could be given. D- ^{18}O relationships in vegetated soils have not been found in the literature.

Yearly mean of percolate from 3 vegetated lysimeters in the study by Eichler were enriched in deuterium corresponding to ^{18}O enrichments ranging from about -0.5‰ (depletion) to 2‰.¹

From experiments with artificially recharged grass-covered lysimeters in France, Blavoux and Siwertz (1971) concluded that a considerable enrichment in ^{18}O took place, but the details given are too few to allow quantitative conclusions to be drawn.

Allison (1982) presented results from a laboratory experiment where evaporation during 8 days from a 1 m sand column, originally at field capacity, caused enrichment to penetrate to 0.10 m, with $\Delta\delta^{18}O$ of about +5‰ at 0.08 m depth and +15‰ at 0.03 m (from Fig.2 in Allison, 1982).

6.2.4 Groundwater

The $\delta^{18}O$ of groundwater is often found to agree closely with mean $\delta^{18}O$ of precipitation. Because of the seasonal variation of groundwater recharge and the effect of soil water storage, this agreement does not, however, exclude the possibility of enrichment of the precipitated water during the recharge as discussed in the preceding sections. Some comments are made below on isotopic changes within the groundwater zone.

For groundwater lying deeper than a few metres below soil surface, the only source of isotopic change would be isotopic

¹ ^{18}O enrichment is estimated assuming a slope of the D- ^{18}O line of 5 because of kinetic effects during evaporation.

exchange with the rock material. This process, generally resulting in ^{18}O enrichment of water, is of significance only in thermal waters (Truesdell and Hulston 1980).

Groundwater close to the soil surface may be subjected to transpiration and direct evaporation, including exchange with the atmosphere. Of these processes only the latter ones are fractionative, as discussed above. In contrast to conditions within the unsaturated zone, it is very unlikely that the bulk volume of groundwater will be affected by fractionation processes at the surface. The reason is that very shallow groundwater mainly occurs in discharge areas for groundwater where the flow has an upward component, preventing the penetration of surface enrichment by diffusion and never allowing penetration by advection.

It can be concluded that no isotopic changes are to be expected within the saturated zone during normal conditions, a fact that is an unspoken premise in almost all hydrogeological studies using stable isotopes.

6.2.5. Surface waters in discharge areas

When performing hydrograph separations by ^{18}O , the observed difference between meltwater or rainwater and streamwater $\delta^{18}\text{O}$ is interpreted as a result of mixing of rain- or meltwater with isotopically different groundwater. The aim of the estimates in this section is to investigate to what extent the observed difference could have been affected by enrichment of the surface water during the runoff event, thereby introducing an error in the hydrograph separations.

In this and the following section, data from field situations from the present investigation are used in more detail than in the foregoing, more general, discussions. The various runoff events are referred to by basin names and year (for snowmelt events) or letters (for rainfall events). The events are defined in the diagrams in Appendix 1, Figs.A1.1-9, showing isotope and discharge data for the periods of measurement.

This section is rather speculative, since complete field data are seldom available for use as input to the applied enrichment models, and even more so since there are very few, if any, isotopic data available for checking the validity of the calculated enrichments.

Wet areas connected to the stream and the stream itself

The simplest way to estimate the enrichment of surface water during a runoff event is to estimate the total net isotopic flux from the atmosphere to the wet surfaces in the basin and calculate how large is the total enrichment caused by this flux on the total discharged volume. It is assumed that all surface reservoirs appear and disappear during the runoff event, and that water losses by evaporation from the wet area can be disregarded from the point of view of water balance. With these assumptions, the volume of water subjected to enrichment equals the discharged volume. The problem is equivalent to estimating the enrichment of a water body without inflow or outflow, and with an area equal to the wet area and a volume equal to total discharged volume. By analogy with Eq.(6.13) we get the mean enrichment $\Delta\delta_s$ of the water as

$$\frac{\Delta\delta_s}{\Delta t} = \frac{E(\delta_s - \delta_E)}{\frac{\bar{Q} \cdot \Delta t}{AY}} \quad (6.21a)$$

where Δt = duration of the event

E = mean evaporation rate from wet areas

$\bar{Q}$ = mean discharge during the event

A = basin area

Y = basin fraction of wet area during the event

Introducing the specific discharge, the enrichment is given by

$$\Delta\delta_s = \frac{E(\delta_s - \delta_E)}{\bar{q}} Y \quad (6.21b)$$

where $\bar{q}$ = mean specific discharge (discharge per total basin area)

At relative humidities less than, say, 75% a value of $\delta_s - \delta_E$ can be obtained from Fig.6.4. At higher relative humidities Eq.(6.21) cannot be used, since $\delta_s - \delta_E$ is too sensitive to $\delta_s - \delta_a$, which changes during the enrichment. Enrichment of a water body then has to be estimated in the usual way by Eqs.(6.12) and (6.5) using short time steps.

Nonlinearity in $d\delta_s/dt$ makes Eq.(6.21) overestimate the enrichment since $\delta_s - \delta_E$ in reality decreases with time as δ_s approaches the "steady-state value" (Fig.6.3). For calculated

enrichments of a few per mil the effect of the time averaging done in Eq.(6.21) is small, however. The nonlinearity may also cause an overestimation due to areal variability of δ_s . For constant E, the enrichment will be largest in shallow areas having little throughflow. These more enriched areas will act as barriers to further enrichment of the total flow volume. This latter effect, which is difficult to quantify, is probably also small. It should be noted, however, that the enrichments calculated by Eq.(6.21) constitute upper limits.

An attempt is made below to estimate the enrichment during the spring flood in Nåsten 1979 (cf. hydrograph Fig.A1.5). All surface water is assumed to be meltwater, having an original $\delta_s = -15\text{‰}$. Other variables are:

$$\begin{aligned}\delta_a &= -20 \text{ ‰} \\ q &= 2 \text{ mm/day} \\ PE &= 1 \text{ mm/day} \\ Y &= 0.15\end{aligned}$$

One difficulty is to get a reasonable value for E, since the evaporation rate from wet surfaces probably varies greatly depending on the exposure to the atmosphere. The following evaporation rates are assumed:

Open sites with free water surface (basin fraction Y_1)	$E = PE$ (potential evaporation)
Free water on the forest floor (Y_2)	$E = 0.1 PE$
Wet areas covered by snow (Y_3)	$E = 0$

The figure for the mean fraction of wet area of the basin, Y, is taken from a field survey at maximum flow (Section 8.1.6) giving $Y_{\max} = 0.23$. The field survey further gave $Y_1 = 0.04$, $Y_2 = 0.09$, and $Y_3 = 0.10$, making $E \approx 0.2 PE = 0.2 \text{ mm/day}$. From Fig.6.4, $\delta_s - \delta_E$ is estimated to be $30 \pm 5\text{‰}$. Eq.(6.21b) then gives

$$\Delta\delta_s = \frac{0.2(30 \pm 5)}{2} 0.15 = 0.5 \pm 0.1\text{‰}$$

considering only the accuracy in $\delta_s - \delta_E$. Estimates for Stormyra give mean enrichments of similar magnitudes. Since the difference between meltwater and streamwater is around $2\text{--}4\text{‰}$, an enrichment of 0.5‰ cannot be disregarded.

From the above exercise it is concluded that there may be an enrichment of surface water during snowmelt, which affects the hydrograph separation by Eq.(3.3). The existence of such enrichment could be detected by studies of the $D/^{18}O$ relationship (cf. Section 3.2), but deuterium analyses were not possible in the present study. For want of such data, the above estimates are regarded as pointing out an enrichment possibility, not as practically applicable results.

Considerable enrichment of rainwater is less probable for runoff events caused by rainfall, since the rainwater probably is near isotopic equilibrium with a saturated atmosphere. However, the water running off after the cessation of rainfall is subject to similar enrichment possibilities as during snowmelt, though with somewhat smaller $\delta_S - \delta_E$, since $R_p \approx \alpha R_a$.

Enrichment on wet areas and in the stream will cause overestimation of the groundwater fraction calculated by Eq.(3.3) for snowmelt periods, since the $\delta^{18}O$ of meltwater is lower than that of groundwater. During rainfall events, the enrichment may either lead to an overestimate or to an underestimate of the groundwater fraction, depending on whether rainfall $\delta^{18}O$ is smaller or larger than groundwater $\delta^{18}O$. See further discussion in Section 9.2.

Stream channel flow and surface water reservoirs

Eq.(6.21) can be used for estimates of the enrichment during various parts of the runoff process, for instance, during stream channel flow and flow through occasional surface water reservoirs, provided that the relative humidity is not too high and that the enrichment does not exceed a few per mil. A more general treatment of these two cases is possible by using Eqs.(6.12) and (6.5) without the approximation introduced in Eq.(6.21). This is done below for stream-channel flow, while the enrichment in flooded areas is estimated from Eq.(6.21).

The enrichment of a stream segment flowing from the "source" to the measurement station during continuous inflow from the bordering land (surface or subsurface inflow) is estimated by Eqs.(6.12) and (6.5). Stationary conditions are assumed, with velocity, v , and inflow per unit length of stream, q' , being constant along the stream. The stream cross section is further assumed rectangular with a constant ratio (c) of width (W) to depth (H).

For enrichment estimates of the stream segment by Eqs.(6.12) and (6.5), we need to calculate the variation with time of its surface area, A_h , and its depth.

From the assumptions on flow and channel geometry we have

$$Q = x q' = v t q' \quad (6.22)$$

and

$$W = c H \quad (6.23)$$

where Q is stream discharge at distance x from the source, v is velocity and t is time of flow from the source. The vertical cross section, A_v , is obtained in two ways as

$$A_v = \frac{Q}{v} \quad (6.24)$$

and

$$A_v = W H = c H^2 \quad (6.25)$$

Eqs.(6.22), (6.24), and (6.25) give

$$H = \sqrt{\frac{q' t}{c}} \quad (6.26)$$

The surface area, A_h , of a segment of length l is

$$A_h = W l = H c l = l \sqrt{q' t c} \quad (6.27)$$

Eq.(6.12) can now be applied to a reservoir having

$$I = \text{inflow per unit reservoir area} = \frac{q' l}{A_h} = \sqrt{\frac{q'}{t c}}$$

$$F = \text{outflow} \quad = 0$$

$$S = \text{reservoir depth} \quad = \sqrt{\frac{q' t}{c}}$$

Enrichment in the channel from the source to the discharge station is estimated with the above model for one of the larger basins in the present study, Stormyra, with a basin area of 4 km^2 . The simulations are performed for a 6 km stream, roughly corresponding to the longest stream reach in the basin. The situations are chosen to represent streamflow during snowmelt

and after the cessation of rainfall, with enrichment variables according to Table 6.3.

Table 6.3 Hypothetical enrichment variables for estimation of enrichment in the stream channel, Stormyra basin.

	summer	spring
Evaporation, E (mm day ⁻¹)	4	1
Relative humidity, h (%)	70	70
$\delta^{18}\text{O}$ of the atmosphere, δ_a (‰)	-18	-18 ^a
$\delta^{18}\text{O}$ of the inflow, δ_I (‰)	-12	-18 ^b

^a A high value is chosen to establish an upper limit of enrichment.

^b Assumed to be only meltwater, gives maximum enrichment.

The results are presented in Fig.6.8, showing the enrichment $\Delta\delta = \delta_{\text{out}} - \delta_{\text{in}}$ as a function of discharge at the outlet (Q), velocity (v), and ratio of width to mean depth (W/H).

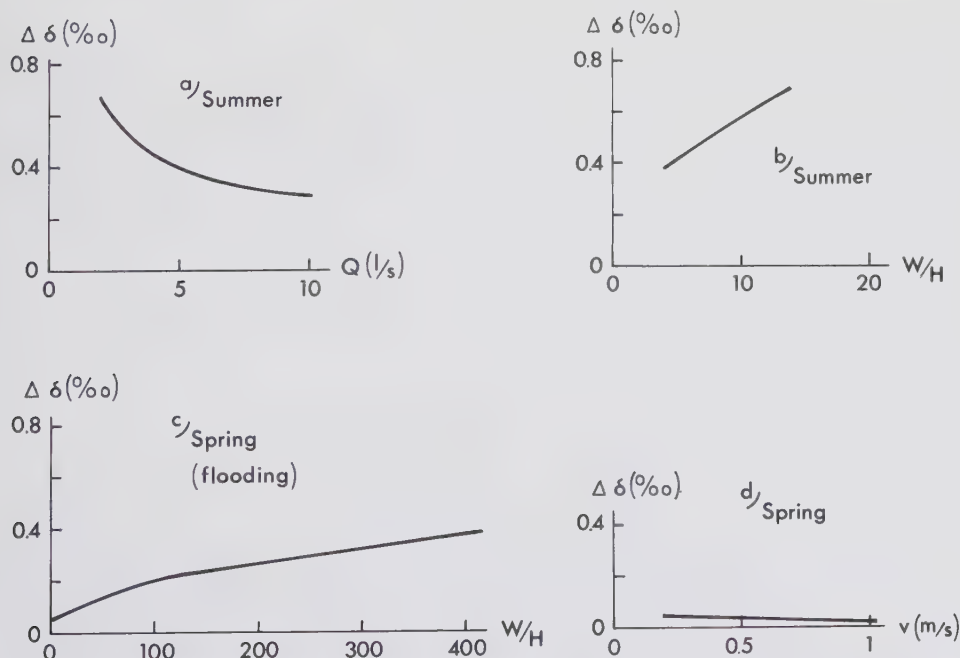


Fig.6.8 Enrichment ($\Delta\delta$) of streamwater during flow through a basin, calculated using data assumed to represent the conditions in the Stormyra basin. Q = stream discharge at the basin outlet, W = stream width, H = stream depth, v = velocity of the water.

For the enrichments shown in the figure, the calculated mean width along the stream is about 0.5-1.0 m for periods without flooding, and up to around 10 m for periods with flooding.

With the above geometrical model (Eqs.6.22 - 6.27) the stream area can be estimated from velocity and discharge. If this estimated area is used, very similar values of $\Delta\delta$ are obtained by Eq.(6.21) from one time step only.

A more realistic case than flooding ($W/H > 100$) along the whole reach of the stream is water flow through certain flooded areas at the lower part of the basin. In Stormyra, flooding of say, a $5 \cdot 10^4 \text{ m}^2$ meadow area just above the discharge station commonly occurs at high flows ($Q > 100 \text{ ls}^{-1}$). Using these enrichment variables, the enrichment during flow of 100 ls^{-1} through the flooded area is estimated by Eq.(6.21), giving $\Delta\delta = 0.8\text{‰}$ during "summer" and 0.3‰ during "spring".

6.2.6 Snow

Performance of hydrograph separation by ^{18}O during snowmelt or modelling of the ^{18}O flow through a basin requires a thorough knowledge of the $\delta^{18}\text{O}$ of meltwater. This entity is preferably measured directly, by use of a snow lysimeter (see construction in Herrmann 1978). In the absence of such measurements, less accurate values of meltwater $\delta^{18}\text{O}$ can be obtained from measured $\delta^{18}\text{O}$ of winter precipitation and snowpack samplings. This latter procedure is applied at the present snowmelt hydrograph separations, since no systematic lysimeter measurements were undertaken.

Important questions to be answered are:

1. What is the relationship between $\delta^{18}\text{O}$ of snowpack at a certain moment and that of meltwater?
2. What is the relationship between $\delta^{18}\text{O}$ of snowfall and that of snowpack, and how does $\delta^{18}\text{O}$ of snowpack change with time?
3. How is the areal distribution of snowpack $\delta^{18}\text{O}$ within a basin?

These questions are briefly discussed in this section starting from results reported in the literature and partly illustrated by the present model for isotopic changes. Fractionation processes in a snowpack are treated in more detail in Section

7.2 in attempts to explain the observed isotopic changes of snowpacks and to calculate the $\delta^{18}\text{O}$ of meltwater from these changes.

The use of isotopes in snow and ice has been reviewed by Moser and Stichler (1980) and by Arnason (1981).

Stable isotopes in snow have most often been studied in glaciated areas, i.e., in snowpacks not melting completely during spring and summer. The glacier-related studies have played an important role in enhancing the present-day knowledge of stable isotopes of water (Epstein and Sharp 1959; Dansgaard, 1961).

Examples of isotope studies in seasonal snowpacks are those by Martinec et al. (1977) and Herrmann and Stichler (1978) made in high and rand-alpine areas in central Europe, respectively, and those by Judy et al. (1970) and Krouse and Smith (1973) treating snowpacks in the United States.

Isotopic changes within a snowpack are caused by two different processes: by molecular exchange with the atmosphere and by fractionation during melting and the subsequent drainage of meltwater. The former processes could be described by the models discussed in Section 6.1, modified to account for the incomplete mixing in the snowpack. Fractionation during melting-freezing is governed by the corresponding fractionation factor

$$\alpha_{s \rightarrow l} = \frac{R_s}{R_l} \quad (6.28)$$

where $\alpha_{s \rightarrow l}$ = fractionation factor for solid-liquid

R_s = isotope ratio of solid (ice)

R_l = isotope ratio of liquid (water)

The ^{18}O fractionation factor for the solid-liquid transition is considerably smaller than that for liquid-vapour transition, the former being 1.003 (O'Neill 1968) as compared to 1.012, both factors given for 0°C. The fractionation during melting-freezing can be interpreted as if the heavy molecules have a higher freezing temperature than the lighter ones, making the heavy molecules more inclined to freeze and less so to melt. From the definition of δ it follows that liquid water in equilibrium with ice is depleted in ^{18}O by about 3‰.

Fractionation during melting-freezing is often restricted by incomplete mixing. Fractionation during melting of solid ice is completely prevented due to the very slow molecular diffusion in ice. The water molecules of the meltwater are identical to those of the melted ice layer. Also, fractionation during freezing may be restricted in many natural situations because of slow molecular diffusion in water and the absence of turbulent mixing in freezing situations.

Relationship between snowpack and meltwater $\delta^{18}\text{O}$

During melting of a snowpack, the original melting is probably non-fractionative, but as the meltwater percolates through the snowpack, melting-freezing will tend to establish isotopic equilibrium between water and ice. The effect depends on the time of contact between snow and water, and also on the isotopic layering of the snowpack. Meltwater depleted in stable isotopes as compared to the snowpack was measured in laboratory experiments by Arnason (1969) and Herrman et al. (1981). Búason (1972) derived a theoretical model to calculate the fractionation during snowmelt or rainwater drainage in a snowpack. The model was tested under natural conditions with good results in a lysimeter experiment reported by Arnason et al. (1973), in which depleted meltwater was measured. Depleted meltwater was further reported from lysimeter experiments by Martinec et al. (1977), who also showed a few cases of enriched meltwater from an isotopically strongly layered snowpack. Stichler and Herrmann (1977) reported meltwater depleted in deuterium corresponding to a depletion in ^{18}O of up to 1.8‰ from 14 lysimeter samplings during a melting period of three weeks. This latter experiment was conducted on a snowpack of a thickness comparable to those encountered in the present investigation. The authors stressed the fact that the difference between snowpack and meltwater δ -values diminishes as the intensity of meltwater production increases. This finding can be explained by different times of contact between meltwater and snow. A graphic example of such a relationship is given in Herrmann and Stichler (1978).

The isotopic composition of a snowpack

The isotopic composition of a snowpack reflects the isotopic composition of the snowfalls earlier during the winter. Since the δ -values within and between precipitation events are highly variable, the snowpack is layered with respect to ^{18}O , even if

diffusional effects tend to smooth differences in isotopic content (Judy et al. 1970). Good agreement between the isotopic content of a snowpack and that of the preceding snowfalls was reported by Moser and Stichler (1975) among others. The effect of meltwater percolation and the related fractionation processes is to homogenize the snowpack with respect to stable isotopes, as demonstrated by the results of Martinec (1974).

A common observation is the gradual enrichment of the snowpack occurring during the ageing of snow, both during the accumulation period and, more pronouncedly, during the melting period. Enrichment of the top layers at temperatures below freezing point was reported by Moser and Stichler (1970) and by Martinec et al. (1977), who both ascribed the enrichment to evaporation and to processes taking place during the metamorphosis of the snowpack. Examples of depletion of thin surface layers (1-2 cm), ascribed to condensation, were reported by Moser and Stichler (1975) and by Herrmann et al. (1978). In laboratory experiments Moser and Stichler found ^{18}O enrichment in snow samples undergoing evaporation of 0.2‰ per % evapo-rated mass of the snowpack.

From isotope balance of snow in a lysimeter, Stichler et al. (1981) estimated evaporation from the snowpack during the melting period using the fractionation model by Craig and Gordon. Reasonable agreement was achieved between evaporation calculated from isotope balance and that calculated from mass balance.

Possible mechanisms of snowpack enrichment by the atmosphere

As discussed above, evaporation from solid ice is non-fractionating due to the very low rate of molecular diffusion in ice, making the isotope content of the evaporate identical to that of the evaporated layer. A snowpack, being built by ice crystals, could be expected for the same reason to show no effects of fractionation by evaporation. But, as demonstrated by many studies, considerable enrichment may take place in a snowpack under non-melting conditions.

Let us discuss the mechanisms of snowpack enrichment by applying the model for isotopic change of a well mixed water body to a snowpack. Nāsten, February 1-15 1980, can be chosen as an example. With the observed relative humidity (90%) and a $\delta^{18}\text{O}$ of the atmosphere of -22‰, in accordance with later

observations in the same area by Saxena (1987), an enrichment rate of 4‰ month^{-1} is obtained for an evaporation of 0.1 mm day^{-1} . The calculated rate of enrichment is similar to the measured value, about 3‰ month^{-1} (cf. Section 7.2, Fig.7.3). Although the enrichment variables have been chosen rather arbitrarily, this agreement suggests the presence of some process which compensates for the incomplete mixing within the snowpack.

Vapour pressure differences between different parts of snow crystals may cause isotopic exchange between snow crystals and the air trapped within the snowpack. The vapour pressure is greater above needles and small ice crystals than above plane surfaces, causing recrystallization by evaporation from needles and small crystals, and condensation on plane surfaces and large crystals. This metamorphosis of snow proceeds continuously throughout the winter. Due to this process there may be a greater molecular exchange between atmospheric vapour and snow than would be expected from the vapour pressures of the air and the snow surface. If the recrystallization takes place in a closed space, no isotopic change will occur. But if the vapour is mixed with atmospheric vapour on its way from evaporation to condensation, the process is likely to cause isotopic changes in the snowpack.

From isotope-balance considerations for the process of recrystallization it can be shown that with $\delta_{\text{snow}} - \delta_a = 10\text{‰}$, a mass rate of recrystallization (evaporation and subsequent condensation per unit mass) of 0.22 month^{-1} is needed to explain an enrichment rate of 1‰ month^{-1} . According to experiments by Wakahama (1968) the rate of recrystallization seems to be well in excess of this value. Recrystallization within the snowpack could thus be an important factor in the enrichment process of a snowpack. The process presumes a considerable gas exchange between the snowpack air and the atmosphere.

Even if estimates of enrichment rates in snowpacks by Eqs.(6.12) and (6.5) or by Eq.(6.16) are uncertain due to restricted mixing in the snowpack, information about conditions for enrichment/depletion can be obtained from Eq.(6.16) putting $d\delta_s/dt = 0$. The curve in Fig.6.9 gives the difference between $\delta^{18}\text{O}$ of snow and atmosphere at which no isotopic change takes place, as function of the ratio between the vapour pressures of

the snow surface and the atmosphere, e_s/e_a .¹ The curve is equivalent to the curve for "steady-state value" of isotopic change, calculated with α -values for surface waters and presented in a slightly different way in Fig.6.3. The areas separated by the curve characterize conditions for depletion and enrichment respectively.

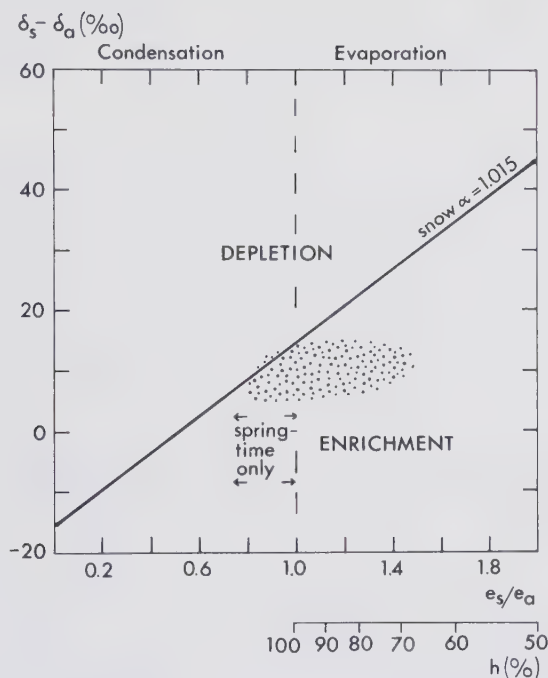


Fig.6.9 Conditions for ^{18}O enrichment or depletion of snow by molecular exchange with the atmosphere. δ_s and δ_a are the $\delta^{18}\text{O}$ of the snow and atmospheric vapour respectively, e_s and e_a the corresponding vapour pressures, and h the relative humidity with respect to the snow surface. The dotted area symbolizes the combinations of $\delta_s - \delta_a$ and e_s/e_a to be expected.

From $\delta^{18}\text{O}$ values of the atmospheric vapour (δ_a) in Uppsala, Sweden, reported by Saxena (1987), and the corresponding values of the snowpack (δ_s) observed in the Nåsten basin, close to Uppsala, values of $\delta_s - \delta_a$ around $5\text{--}10\text{‰}$ seem reasonable.² It is seen from Fig.6.9 that relative humidities less than 100%, i.e., evaporation, in general give enrichment as expected. A

¹ $\alpha_{\text{ice-vapour}} = 1.015 = \alpha_{\text{ice-liquid}} \cdot \alpha_{\text{liquid-vapour}}$ at 0°C

² This difference, which also was found by Saxena during single snowfalls, is considerably smaller than the difference at isotopic equilibrium between the snow and a saturated atmosphere, 15‰ .

somewhat surprising finding from Fig.6.9, however, is that enrichment may well occur also during condensation. A snow cover differs from a shallow water body in that condensation may prevail for comparatively long periods (several days). The situation occurs frequently in springtime when warm, moist air enters a snow-covered area, where the surface temperature can never exceed 0°C . The warm air is likely to have comparatively high δ -value, making $\delta_{\text{snowpack}} - \delta_a$ small, say 10‰ . At moderate supersaturation with respect to the snow, this will cause enrichment according to the figure.

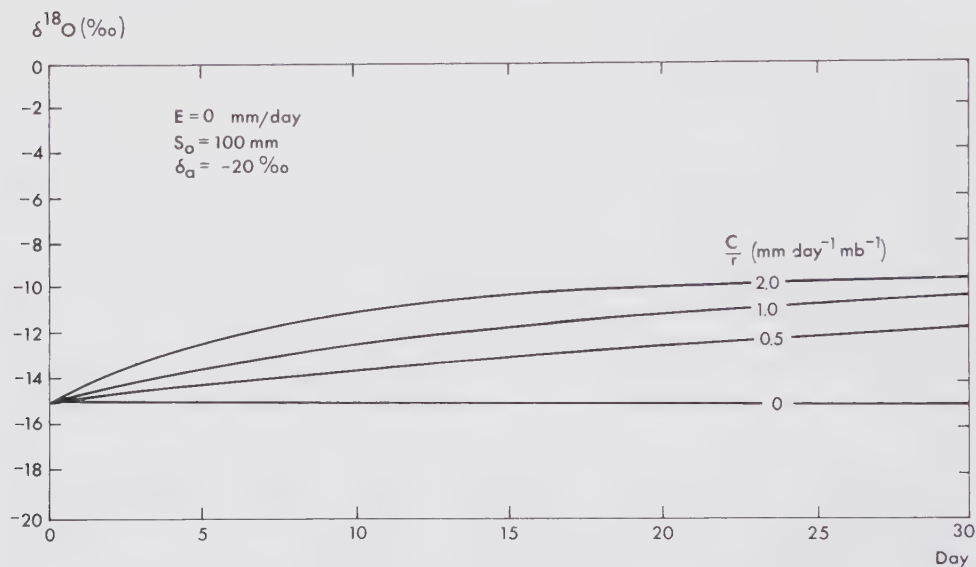


Fig.6.10 Calculated enrichment of snow at 100% relative humidity for various rates of turbulent exchange (described by the transfer coefficient C/r , see text). E = evaporation, S_0 = water equivalent of the snowpack, $\delta_a = \delta^{18}\text{O}$ of the atmospheric vapour.

Possible rates of enrichment during a period of no evaporation ($e_s = e_a$) are illustrated by applying Eq.(6.16) to a "well-mixed" snowpack of 100 mm water equivalent. Fig.6.10 shows calculated isotopic changes for different rates of turbulent exchange with δ_a being constant. The values of turbulent exchange, C/r , have been chosen to give reasonable evaporation rates in unsaturated conditions. $E = 0.5$ mm day⁻¹ at $e_s = 6$ mb (0°C) and $h = 70\%$ gives $C/r = 0.3$ mm day⁻¹ mb⁻¹. It should be noted that the assumption of complete mixing is not realistic, but, as shown by the example from N  sten, the rate of enrichment of a snowpack still seems to be fairly well described by the enrichment model.

Areal variability of snowpack $\delta^{18}\text{O}$

Few results are available from the literature on the areal variability of snowpack δ -values. Meiman et al. (1973) reported ranges in the deuterium content of the snowpack in a regional study ($\approx 10^4 \text{ km}^2$, 24 samples) in Colorado, western United States, corresponding to $\delta^{18}\text{O}$ ranges of 2‰ and of $0.8\text{--}1.6\text{‰}$ in two basins ($32 \text{ km}^2/9$ samples and $9 \text{ km}^2/6$ samples, respectively).

Moser and Stichler (1975) present results from studies of altitude dependence of snowpack δ -values, showing that the altitude dependence of precipitation $\delta^{18}\text{O}$ ($0.3\text{‰} (100\text{m})^{-1}$) is likely to be found only in fresh snow. Data showing the opposite trend with the altitude, i.e., increased snowpack δ -value by altitude, are shown for old snowpacks. The authors ascribed these findings to different rates of enrichment during the ageing of the snow at various sites.

6.2.7 Conclusions on enrichment during water flow through a basin

Rainfall interception

The throughfall may be enriched in ^{18}O by a few tenths of permil as compared to the precipitation. This enrichment is of little importance when performing isotopic hydrograph separation (cf. the uncertainty discussion, Chapter 9). It may, however, introduce a small systematic error when ^{18}O is used for modelling the water flow through a basin.

Infiltration

During a rainfall, enrichment of infiltrating water (ponded or non-ponded) is unlikely, since the water and the saturated air are close to isotopic equilibrium. For the same reason, no enrichment is expected if Hortonian overland flow were to occur during the rainfall. After a rainfall, there could be a considerable enrichment of water infiltrating from a ponded area into a soil of low infiltration capacity. But since the infiltration capacity of till soil is generally great, few ponds appear in recharge areas. If a pond does remain after a very heavy rainfall, the enrichment would be small due to a high infiltration capacity.

For infiltrating meltwater the atmospheric conditions are more favourable to enrichment. But evaporation and molecular exchange with the atmosphere are effectively prevented for water infiltrating under a snowpack, making enrichment unlikely. The same applies to Hortonian overland flow under a snow cover. The infiltration situation which is most favourably disposed to enrichment is ponding meltwater exposed to the atmosphere. However, estimates made using infiltration capacities expected in till soils show that the mean enrichment of the infiltrating water is unlikely to exceed a few tenths of per mil.

Soil water

Due to the seasonal variations in the isotopic composition of precipitation and of the water loss by evapotranspiration, comparisons between the $\delta^{18}\text{O}$ of soil water or percolate and the precipitation give no direct information on enrichment during infiltration and in the unsaturated zone.

In bare soils undergoing evaporation, enrichment takes place in the top few decimetres. In humid areas, the surface enrichment is transported downwards mainly by downward water flow. Observations of enrichment of deuterium in the unsaturated zone corresponding to a 2‰ enrichment in ^{18}O have been reported.

Transpiration is probably non-fractionative. No conclusive evidence of enrichment of soil water in vegetated areas has been found in the literature.

Surface waters in discharge areas

The mean enrichment of the water discharged during a runoff event can be estimated from the total net isotopic flux from the atmosphere to the wet areas, including the stream itself. Estimates for a spring flood in the Nåsten basin show that mean enrichment of several tenths of per mil is possible. The estimates are, however, very uncertain since they involve guesses of the evaporation rates from wet areas outside the stream.

Enrichment during stream channel flow through reservoirs is somewhat less uncertain, since evaporation is better known. From estimates for one of the present study basins it can be concluded that:

Streamflow enrichment is negligible ($<0.1\text{‰}$) at moderate or high flows, during both spring and summer, as long as the water is flowing in the channel only.

Streamflow enrichment of $0.5\text{--}1.0\text{‰}$ during low flow in summer is probable.

Enrichment of $0.5\text{--}1.0\text{‰}$ may well occur during flow through large flooded areas.

The above comments on enrichment during streamflow and flooding apply to the two largest basins of the study, Nåsten and Stormyra. In the other basins the enrichment is less, since they are steeper and less disposed to flooding.

Snow

A considerable enrichment of the precipitated water (snow) takes place during snow storage. This enrichment takes place in recharge areas as well as in discharge areas for groundwater. The fractionation during melting-freezing tends to make the meltwater depleted in heavy isotopes and thus to enrich the remaining snow. This fractionation mainly affects the time distribution of isotopic input to the soil or streams and less so the total isotopic input. The fractionation caused by exchange with the atmosphere, on the other hand, increases the total input of heavy isotopes to the basin.

7 ISOTOPE CONTENT OF THE WATER INPUT

As a background to the interpretation of streamwater $\delta^{18}\text{O}$ in Chapter 8, isotopic data on the water input to the basins, i.e., on precipitation and snowpack, are presented in this chapter. The data on precipitation $\delta^{18}\text{O}$ are presented briefly, whereas the data on snowpack $\delta^{18}\text{O}$ are analysed in some detail in order to make it possible to generalize the results.

Precipitation amount and $\delta^{18}\text{O}$, streamwater $\delta^{18}\text{O}$, and stream discharge are shown for each investigated basin and year in Appendix 1, Figs.A1.1-9.

7.1 Precipitation

Hydrograph separation of rainfall-generated runoff events by ^{18}O requires a basin mean value of rainfall $\delta^{18}\text{O}$. In the present study, event mean values of rainfall $\delta^{18}\text{O}$ from one station, within or close to the basins, are used. To give an idea of the accuracy of such a procedure, a few data on areal and temporal variations in precipitation $\delta^{18}\text{O}$ are commented upon in this section. The possibility of estimating daily values of precipitation $\delta^{18}\text{O}$ from meteorological data is also discussed in view of the present observations.

7.1.1 Seasonal variation

The time courses of precipitation $\delta^{18}\text{O}$ (Figs A1.1-9) reveal a large scattering in the daily values. Seasonal variation is seen in the monthly mean values (not shown here), with low values in winter and high values in summer. During the years investigated, the seasonal variation was least pronounced in Gårdsjön. This agrees with the findings on longer time series from Swedish stations by Burgman et al. (1987); cf. Section 3.2.2.

Hydrograph separation by means of ^{18}O relies on the damping of input variations in $\delta^{18}\text{O}$ that takes place during water flow through a basin. Great seasonal variation in precipitation thus favours the method. But because of the large scattering of the daily values, hydrograph separation of some runoff events can be performed also when the seasonal variation is small or absent.

7.1.2 Variations within storms

As shown by several studies, e.g. Ambach et al. (1975) (see discussion in Section 3.2.2), there may be large variations of precipitation $\delta^{18}\text{O}$ within storms. In the present study, the variations within a few rainfall events were investigated in Gårdsjön, see Section 8.2.1 Figs.8.9b and 8.10, and in Svartberget (Fig.7.1). These results show that very large in-storm variations may occur (11‰ in Gårdsjön A, Fig.8.9b), but from the limited amount of data available it seems as if such large variations were rare. With the volume resolution in the above examples, usually 1-5 mm per sample, variations of 1 - 2‰ around the weighted mean seem common. Implications for hydrograph separations are discussed in connection with the analysis of intensively sampled rainfall-generated runoff events (cf. Section 8.2.1).

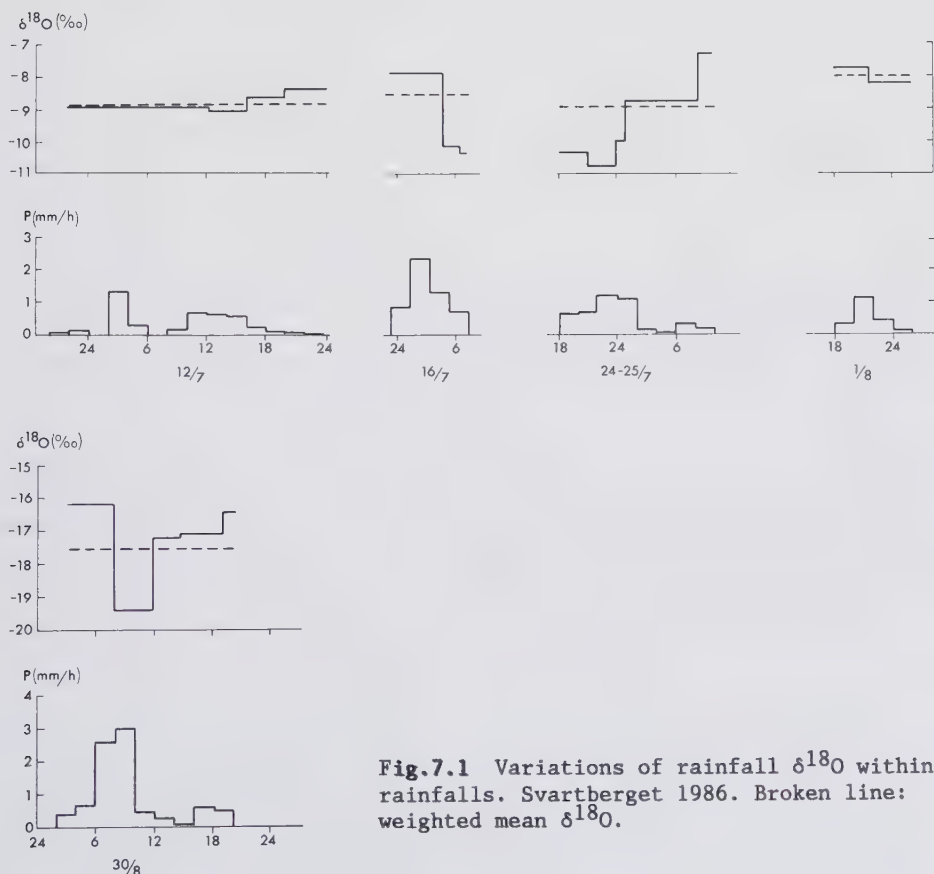


Fig.7.1 Variations of rainfall $\delta^{18}\text{O}$ within rainfalls. Svartberget 1986. Broken line: weighted mean $\delta^{18}\text{O}$.

7.1.3 Areal variability

The areal variability of precipitation $\delta^{18}\text{O}$ within the basins was not studied systematically in this investigation. The snowpacks were sampled at several points within the basins, but these surveys give little information on the areal variability of the snowfall, since differences in $\delta^{18}\text{O}$ between the stations are also the results of different rates of isotopic exchange with the atmosphere and of different rates of melting, cf. Section 7.2. To give an idea of the expected areal variability of precipitation $\delta^{18}\text{O}$, the results of a few indicative precipitation samplings are shown in Table 7.1. The rainfalls and snowfalls were all of frontal origin, having low intensities.

Table 7.1 Examples of areal variability of precipitation $\delta^{18}\text{O}$

Exp. No.	Location	Date	n	Δh (m)	ΔL (km)
1	Gravendal ^a	17/9 1980	5	95	9
2	Finnmossen ^b I	21/9 1981	4	30	3
3	Finnmossen II	21/9 1981	7	1	0.01
4	Uppsala ^c	19/12 1979	7	10	22

n = number of observation

Δh = altitude range

ΔL = maximum distance between stations

^a 50 km SE of Buskbäcken

^b 50 km W of Buskbäcken

^c Near Nåsten

Precipitation amount					$\delta^{18}\text{O}$ of precipitation or snowpack		
Exp. No.	P (mm)	stand. dev. (mm)	range (mm)	rainfall intensity (mm h ⁻¹)	δ_p (‰)	stand. dev. (‰)	range (‰)
1	4.5	0.4	4.0-5.0	0.4	-8.81	0.14	-9.02 - -8.68
2	12.9	1.3	11.3-14.5	1.5	-12.66	0.04	-12.71 - -12.61
3	14.5	0.8	12.6-14.9	1.5	-12.64	0.06	-12.76 - -12.60
4	snowpack, $\approx$ 6 mm				-16.6	0.39	-17.2 - -16.1

In Gravendal and Finnmossen, single rainfalls were collected in forest clearings or glades. At some of the stations at Finnmossen, rainfall was collected at several points (6-10 points), separated by a few metres. For one of these stations, reported as Finnmossen II in Table 7.1, analyses for ^{18}O were performed on all samples. The data used for Finnmossen I in the table are mean values for each station. Except for the intensively measured Finnmossen II, station means of $\delta^{18}\text{O}$ were obtained by mixing equal volumes to form one sample for ^{18}O analysis.

In the Uppsala experiment, the snowpack was sampled in flat agricultural land along a 22 km straight line. At each station, a composite sample was formed out of 4 samples. The snowpack originated from a few precipitation events, 1-6 days before the day of sampling. In view of the cold weather, of the short time elapsed since the formation of snowpack, and of similar exposure to the atmosphere at all stations, the areal variability of these particular snowpack samples is probably well in keeping with that of precipitation.

The accuracy in δ_p needed at hydrograph separation of rainfall-generated runoff events by Eq.(3.3) depends on the magnitude of $|\delta_p - \delta_g|$, cf. Chapter 9. Accuracies around $\pm 0.2 - \pm 1.0\text{‰}$ are considered acceptable for most of the hydrograph separations performed here.

The small-scale experiment (Finnmossen II) shows, as expected, a variability close to the precision of the analysis for ^{18}O . The variability is also very small in Finnmossen I and Gravendal. In the Uppsala experiment, the variability may partly be an effect of different snow storage at the sites.

It is concluded that one sampling point for precipitation would give a sufficient accuracy of δ_p for these precipitation events. It must be noted, however, that convective showers are likely to give larger areal variability in precipitation $\delta^{18}\text{O}$ (cf. Dansgaard, 1961) as well as in precipitation amounts.

7.1.4 Correlation with air temperature

As discussed in Section 3.2.1, the $\delta^{18}\text{O}$ of precipitation has been found to be well correlated with ground air temperature, both for long-term mean values at different sites and for monthly values at single sites (e.g. Yurtsever and Gat 1981). Significant correlation between the two variables has also been

found for daily values at single sites, but this relationship is much weaker.

In attempts to obtain a means to estimate precipitation $\delta^{18}\text{O}$ when measured values are not available, linear regression was performed between daily values of precipitation $\delta^{18}\text{O}$ and daily values of precipitation amount, vapour pressure, and air temperature. Analysis of some of the data showed that the only statistically significant correlation was that to air temperature.

In Table 7.2 are listed results of regression analysis on all daily values of precipitation for which good temperature data are available. The coefficients of correlation are small, but they show significant relationships in all basins, except at Gårdsjön.

Table 7.2 Linear regression between air temperature (T) and $\delta^{18}\text{O}$ of precipitation (δ_p), daily values.

Basin	Period	n	a ‰°C ⁻¹	b ‰	r
Buskbäcken	1980-1982 all year	409	0.36	-13.5	0.65
	1980-1982 summers	109	0.43	-15.0	0.44
Gårdsjön	1980-1981 all year	151	0.14	-9.6	0.29
	1980-1981 summers	58	0.11	-9.6	0.10
Nåsten	1979-1982 all year	256	0.36	-13.3	0.65
	1979-1982 summers	55	0.30	-12.5	0.32
Stormyra	1979-1982 all year	149	0.22	-12.1	0.44
	1979-1982 summers	24	0.63	-19.4	0.42
Svartberget	1981-1982 all year	162	0.41	-13.5	0.67
	1981-1982 summers	33	0.65	-16.9	0.47
All basins	1979-1982 all year	1188	0.36	-13.1	0.64

Regression equation $\delta_p = aT + b$
 n = number of observations
 r = coefficient of correlation

The slopes of the linear regression lines in the basins having significant relationships range between 0.22 and 0.41 when the whole periods are considered. For the whole set of data, including Gårdsjön, the slope is 0.36. This value agrees with the value found by Yurtsever and Gat (1981), describing the variation of $\delta^{18}\text{O}$ in precipitation from mean monthly air temperature at different stations (cf. Section 3.2).

According to Saxena (1987) the slope agrees with an eddy-diffusive transport mechanism for water vapour, rather than a pure advective transport, which would give a slope around $0.7\text{‰ }^{\circ}\text{C}^{-1}$.

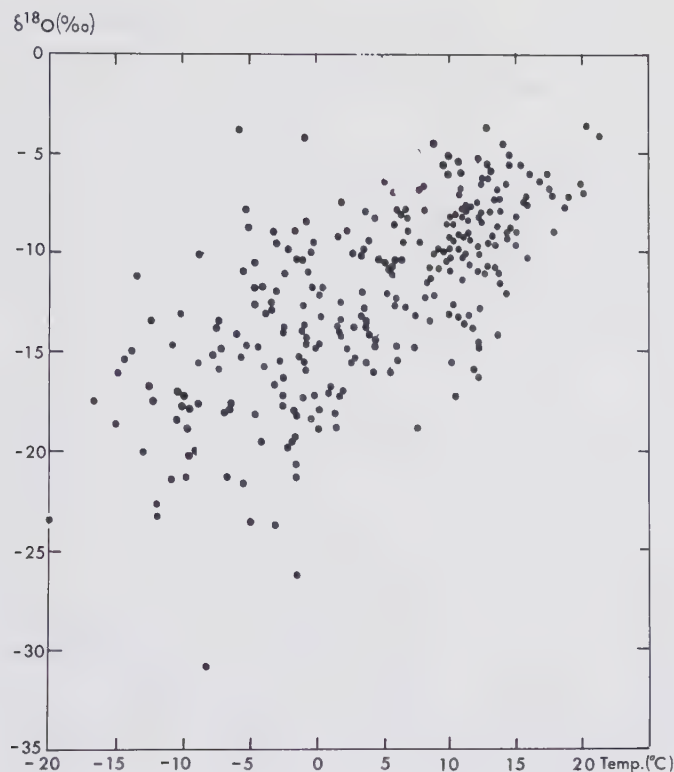


Fig.7.2 $\delta^{18}\text{O}$ of precipitation versus ground air temperature, daily values. Buskbäcken 1980 - 1982.

These regression equations cannot be used for estimates of daily values of precipitation $\delta^{18}\text{O}$. This is seen, for example, from Fig.7.2. For a certain air temperature, the range of δ -values at this station is greater than 10‰ , which is to be compared with the total range of δ -values of about 20‰ . The same conclusion can be drawn from the coefficients of correlation obtained for the summer periods (Table 7.2). By regarding summer values only, the influence of the seasonal variation in air temperature is reduced. The remaining dependence on air temperature is very small, as shown by the small coefficients of correlation.

It is evident that the correlation found for the all-year data to a large degree is a result of the seasonal variations in

precipitation $\delta^{18}\text{O}$ and air temperature. With this in mind, the small coefficients of correlation found at Gårdsjön are natural, considering the small seasonal variation in precipitation $\delta^{18}\text{O}$ in this area.

It is concluded that precipitation $\delta^{18}\text{O}$ cannot be estimated with any acceptable accuracy from air temperature only, neither for hydrograph separation by means of ^{18}O nor for modelling of water flow through a basin by ^{18}O . A few gaps in the daily values of precipitation $\delta^{18}\text{O}$ needed for calculation of the isotope input to snow covers (Section 7.2) and for estimating transit times and reservoir volumes for water in the basins (Section 8.3), are filled, assuming that the daily δ -value equals the observed monthly mean value.

7.2 Snow

In this section, observed changes of snowpack $\delta^{18}\text{O}$ are analysed in an attempt to estimate the $\delta^{18}\text{O}$ of meltwater, which is needed in the hydrograph separations. The snowpacks are looked upon from the point of view of mass and isotope balance, making it possible to quantify the effects of the fractionation processes discussed theoretically in section 6.3.6. Attempts are made to draw generalized conclusions on the relationship between the $\delta^{18}\text{O}$ of precipitation during winter and the $\delta^{18}\text{O}$ of snowpack and meltwater. Knowledge about this relationship is needed when ^{18}O is used for modelling water flow through a basin from data on the $\delta^{18}\text{O}$ of precipitation and streamwater.

Two examples of the temporal variation of the mean $\delta^{18}\text{O}$ and water equivalent of the snowpack during late winter and spring are shown in Fig.7.3, and all such observations are shown in Appendix 2, Fig.A2.1. When calculating the basin mean δ -value, all snow-covered stations have been given the same weight, regardless of water equivalent. The ranges of snowpack $\delta^{18}\text{O}$ found at the surveys are marked in the figures¹. The ranges generally lie in between 1 and 3 ‰.

For most periods, the $\delta^{18}\text{O}$ of the snowpacks increased during late winter and spring, as has also been observed in several other studies (cf. Section 6.2.6). As discussed in Section 6.2.6, the increase is mainly due to molecular exchange with the atmosphere and to drainage of isotopically light meltwater.

¹ The number of stations at each survey varied from 3 to 10, usually more than 6 (cf. Section 5.1.2).

The shallow snowpack in Nåsten 1980 (Fig.7.3) showed a comparatively rapid enrichment, also during non-melting periods. In contrast with most other observed periods, the $\delta^{18}\text{O}$ of the snowpack in Buskbäcken 1981 (Fig.7.3) decreased during some periods. As shown below, these decreases were caused by the supply of isotopically light snowfalls.

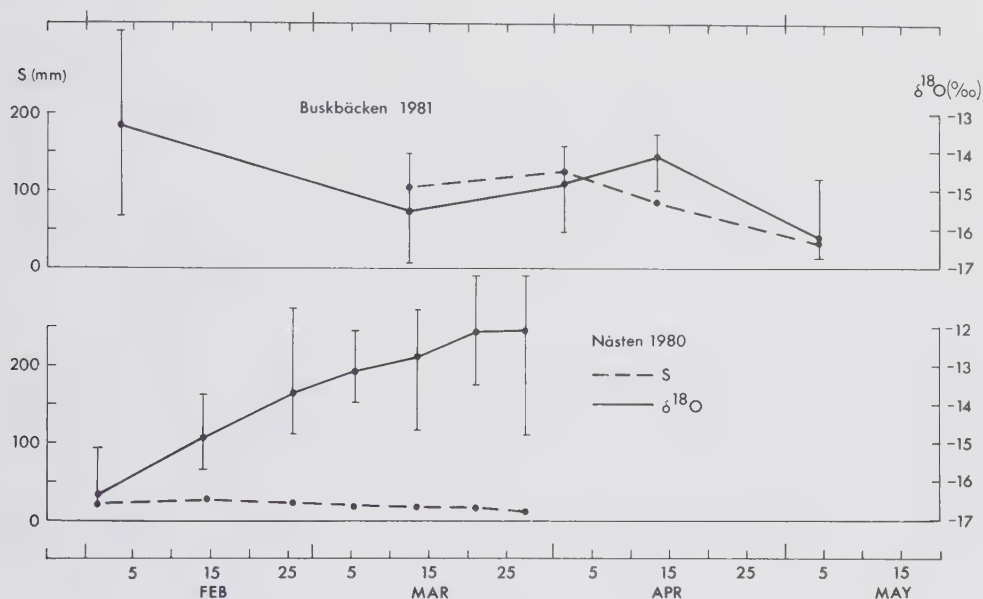


Fig.7.3 Measured $\delta^{18}\text{O}$ and water equivalent (S) of the snowpacks in Buskbäcken 1981 and Nåsten 1980. Ranges of snowpack δ -values shown.

In order to estimate the $\delta^{18}\text{O}$ of meltwater, mass and isotope balances of the snowpacks are analysed below. Although the isotope content of meltwater to a large degree is determined by the vertical isotopic layering of the snowpack, the balance considerations are first performed for the total snowpacks, using the mean $\delta^{18}\text{O}$ obtained from vertical snow cores. This approach is chosen since most of the available data refer to such composite snowpack samples (Figs. 7.3 and A2.1). Mass balance of the total snowpack can be used in estimates of one missing term in the balance, for instance the $\delta^{18}\text{O}$ of meltwater, from measured values of the others. But when predictions of the $\delta^{18}\text{O}$ of meltwater are to be made, the actual isotopic layering of the snowpack has to be considered. In order to visualize the effect of the vertical ^{18}O -layering on the $\delta^{18}\text{O}$ of meltwater, a mass balance based on the isotopic layering of the snowpack is also discussed.

7.2.1 Mass and isotopic balance of a snowpack

Mass and isotope balance of a snowpack extending from time t_1 to t_2 can be written as follows:

$$\text{Mass balance } S_2 - S_1 + M - P = 0 \quad (7.1)$$

$$\text{Isotope balance } S_2\delta_2 - S_1\delta_1 - A\Delta t + M\delta_M - P\delta_p = 0 \quad (7.2)$$

where S = water equivalent of snowpack (mm)

M = sum of meltwater production (mm)

P = sum of precipitation (mm)

$\delta, \delta_M, \delta_p$ = the corresponding isotope contents (‰)

A = net isotopic flux from the atmosphere (‰ mm day⁻¹)

$\Delta t = t_2 - t_1$

subscript 1 and 2 refer to t_1 and t_2 , respectively

Evaporation has been omitted in the mass balance, since evaporation from a snowpack is small compared to the accuracy in determining $S_2 - S_1$ for the periods under consideration. However, the rate of evaporation is present in the factor A in the isotope balance (cf. Section 7.2.4).

Eqs.(7.1) and (7.2) give the isotopic change during the period as

$$\delta_2 - \delta_1 = \frac{P(\delta_p - \delta_1) - M(\delta_M - \delta_1) + A\Delta t}{S_2} \quad (7.3a)$$

or

$$\delta_S = \Delta_p + \Delta_M + \Delta_A \quad (7.3b)$$

where $\delta_S = \delta_2 - \delta_1$ is the total isotopic change

$$\Delta_p = \frac{P(\delta_p - \delta_1)}{S_2} \quad \text{is the isotopic change caused by precipitation}$$

$$\Delta_M = \frac{M(\delta_1 - \delta_M)}{S_2} \quad \text{is the isotopic change caused by meltwater}$$

$$\Delta_A = \frac{A\Delta t}{S_2} \quad \text{is the isotopic change caused by molecular exchange with the atmosphere}$$

Eq.(7.3) can be used to calculate the isotopic composition of meltwater, δ_M , which is needed for the hydrograph separations. The problem is to obtain an accurate value of the net isotopic flux from the atmosphere.

Calculations of the net isotopic flux from the atmosphere

The net isotopic flux from the atmosphere, A , is estimated below from the observed changes of snowpack $\delta^{18}O$ during periods of little or no melting. In later calculations of δ_M , these values of A are extrapolated to melting periods.

The combined effects of meltwater and flux from the atmosphere upon the changes of snowpack $\delta^{18}O$ for the periods shown in Fig.7.3. are seen in Fig.7.4, showing

$$\Delta_M + \Delta_A = \Delta\delta_S - \Delta_P$$

i.e., the measured changes of snowpack $\delta^{18}O$ corrected for changes by precipitation. Corresponding diagrams for all periods are found in Appendix 2, Fig.A.2.2.

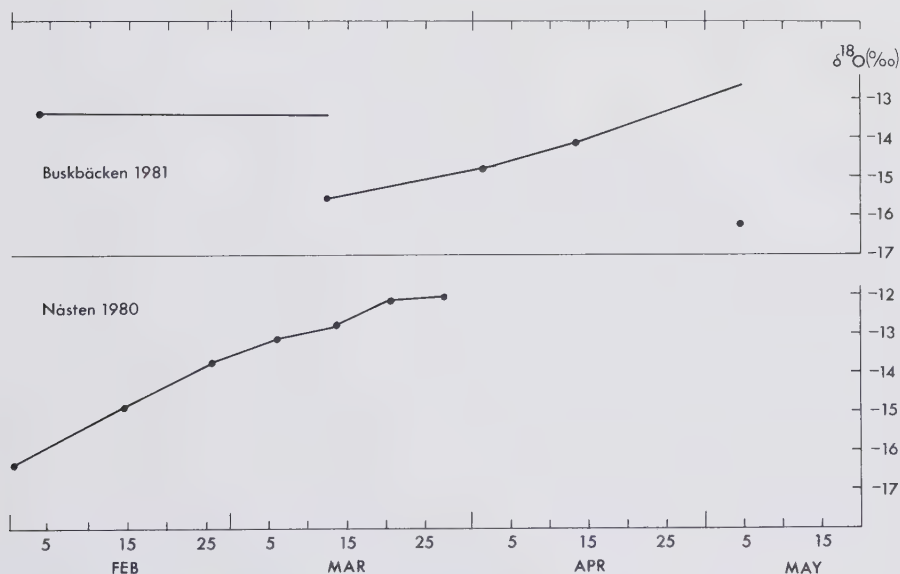


Fig.7.4 Measured $\delta^{18}O$ of snowpacks (dots) and calculated changes of snowpack $\delta^{18}O$ after elimination of the isotopic input by precipitation (lines). Buskbäcken 1981 and Nåsten 1980.

It is seen from Fig.7.4 that snowpack enrichment takes place during both non-melting and melting periods. The rate of enrichment during non-melting is often around 1‰ month^{-1} . A few periods of ^{18}O depletion are also noted.

For periods of little or no melting, A is calculated from

$$\Delta_A = \Delta\delta_S - \Delta_p - \Delta_M$$

and

$$A = \frac{\Delta_A}{\Delta t} S_2 \quad (7.4)$$

Since δ_M is not known, A can be accurately calculated only for non-melting periods. During periods of little melting, however, Δ_M will be a small term. Roughly estimated ranges of $\delta_S - \delta_M$, based on the results of lysimeter experiments, make it possible to estimate ranges of A also for these periods.

Measured isotopic changes, calculated changes by various mechanisms and the resulting net isotopic flux from the atmosphere are listed in Table 7.3. The most accurate values of A are those calculated over long periods having small Δ_p and no melting.¹ For periods containing days with air temperature above 0°C , a value for the upper limit of melting was roughly estimated from $S_2 - S_1$, air temperature, and hydrographs. The value of $\delta_S - \delta_M$ was assumed to fall within $\pm 2\text{‰}$. The calculated values of the net isotopic flux range from -10 to 10‰ mm day^{-1} , with a mean value for all estimates (weighted for period length) of 1.9‰ mm day^{-1} . The values are scattered and no seasonal trends in the respective basins are visible. The negative values encountered during certain periods in Svartberget and Aspåsen, indicating depletion of heavy isotopes by the atmosphere, are somewhat surprising. But the fact that both Svartberget and Aspåsen have negative values at the end of March 1981 suggests that these depletions were caused by a particular weather condition during the period in this part of Sweden and not just results of measurement or sampling errors. No reasons for the comparatively high values of A in Aspåsen have been found.

¹ The initial water equivalent of snowpack, S_1 , has been eliminated in Eq.(7.3) by the mass balance, Eq.(7.1). Due to uncertainty in the terms of Eq.(7.1), particularly in S_1 and S_2 , the mass balance is not fulfilled by the values used during many of the periods for which A has been estimated. For each period, the measured values of P and S_2 have been used. The error in A thus introduced is smaller than would be the case if A were calculated from Eq.(7.2) only.

Table 7.3 Calculation of net ^{18}O flux from the atmosphere to the snowpack

Basin		$\Delta\delta_s$ (‰)	Δ_p (‰)	M (mm)	Δ_A (‰)	S_2 (mm)	A (‰·mm·day ⁻¹)
Aspäsén							
1981	26/3- 2/4	-0.36	0	<10	-0.36±0.11	188	-9.6±3.0
	12/4-17/4	0.43	0.04	<10	0.39±0.15	131	10.2±3.9
	17/4-26/4	0.51	-0.04	<10	0.55±0.13	152	9.3±2.1
	26/4- 1/5	-0.16	-0.39	0	0.23	148	6.8
Σ	12/4- 1/5	0.78	-0.39	<20	1.17±0.27	148	9.2±2.1
Buskbäcken							
1979	13/2- 9/3	1.47	≈0.64	0	0.83	107	5.1
1980	13/3-10/4	0.35	0.32	<10	0.03±0.19	105	0.1±0.7
	16/4-21/4	1.04	0.86	<5	0.18±0.23	42	1.5±1.9
Nåsten							
1980	1/2-15/2	1.43	0	<5	1.43±0.38	26	2.6±0.7
	6/3-14/3	0.34	0.04	0	0.34	18	0.8
	14/3-21/3	0.68	0	<2	0.69±0.24	17	1.7±0.6
	21/3-28/3	0.02	0	<2	0.02±0.25	16	0.0±0.6
Σ	6/3-28/3	1.04	0.04	<4	1.00±0.50	16	0.7±0.4
Stormyra							
1979	8/2- 2/3	0.7	0.7	0	0.7	131	4.2
1980	6/2- 7/3	1.87	0.08	<2	1.87±0.13	31	1.9±0.1
1982	11/2- 2/3	0.55	0.54	<15	0.01±0.22	133	0.1±1.5
Svartberget							
1981	11/3-25/3	0.26	0.04	0	0.22	194	3.0
	25/3- 1/4	-0.14	0	<10	-0.14±0.10	204	-4.1±2.9
	15/4-23/4	0.22	0	<10	0.22±0.13	157	4.2±2.5
	23/4-29/4	0.12	0.02	0	0.10	152	2.4
	29/4- 7/5	0.19	0	0	0.19	154	3.7
Σ	11/3- 1/4	0.12	0.04	<10	0.12±0.10	204	1.2±1.0
Σ	15/4- 7/5	0.53	0.02	<10	0.53±0.13	154	3.7±0.9
1981/82							
	24/11-27/12	-1.61	-2.44	0	0.83	155	3.8
	28/12-21/1	-0.17	0.00	0	-0.17	153	-1.0
	22/1-23/2	0.96	0.38	0	0.58	185	2.1
	24/2-24/3	0.73	0.72	0	0.01	209	0.1
	25/3- 6/4	-0.01	-0.01	<20	0.00±0.22	183	0.0±3.1
	7/4-12/4	0.07	-0.13	<5	-0.06±0.05	207	-2.1±1.7

 $\Delta\delta_s$ = observed change in snowpack $\delta^{18}\text{O}$ Δ_p = isotopic change by precipitation

M = amount of meltwater

 Δ_A = isotopic change by molecular exchange with the atmosphere S_2 = water equivalent of the snowpack at the end of the periodA = net ^{18}O flux from the atmosphere.

In order to make it possible to estimate the net isotopic flux from the atmosphere during melting periods, and to generalize the results, attempts have been made to correlate A to easily available weather variables. From Eq.(6.16) it is seen that the rate of enrichment, and thus the net isotopic flux from the atmosphere, could be expected to be positively correlated to wind speed and air temperature (by e_s and δ_a , effects of which probably dominate over effects of changes in α by temperature) and possibly negatively correlated to e_a . (The latter statement is not certain, since e_a is positively correlated to air temperature.) Multiple linear regression between all calculated values of A and period mean values of the weather variables suggested above gave no significant relationship.¹ One reason might be that δ_a is not univocally related to air temperature, since δ_a is also determined by the origin of the air mass. But probably more important is the influence of the physical structures of the snowpacks on the isotopic exchange between snow and atmosphere, both on the exchange between snow air and the atmosphere, and on the rate of recrystallization (cf. Section 6.2.6).

Since no useful relationship between weather variables and net isotopic flux from the atmosphere could be established, values of A from the non-melting periods were assumed to be valid also for melting periods. The following values of A were chosen for use in estimates of meltwater $\delta^{18}\text{O}$: Aspåsen $8 \pm 3\text{‰}$ mm day⁻¹, the other basins $2.5 \pm 2.5\text{‰}$ mm day⁻¹.

Isotopic composition of meltwater from basin mean snowpack $\delta^{18}\text{O}$

The isotopic composition of meltwater leaving the snowpack is estimated from the isotopic changes of the snowpack by Eq.(7.3b), which can be written

$$\Delta_M = \Delta\delta_S - \Delta_P - \Delta_A$$

$$\text{and } \delta_1 - \delta_M = \frac{\Delta_M}{M} S_2 \quad (7.5)$$

where $M = S_2 - S_1 + P$ (For definition of symbols, see legends to Eqs.7.1-7.3).

Calculated values of the difference between snowpack and meltwater $\delta^{18}\text{O}$ are listed in Table 7.4, which also shows the terms

¹ Complete series of all weather variables were not available for all the basins.

Table 7.4 Calculation of meltwater $\delta^{18}\text{O}$ from changes of snowpack $\delta^{18}\text{O}$.

Basin		$S_2(\delta_2 - \delta_1)$	$P(\delta_p - \delta_1)$	$A \Delta t$	M	$\delta_1 - \delta_M$	$\bar{\delta}_S - \delta_M$
(°/‰)		(°/‰·mm)	(°/‰·mm)	(°/‰·mm)	(mm)		(°/‰)
Aspåsen							
1981	2/4-11/4	156	0	80±30	50	1.5±0.6	2.1±0.6
	1/5- 8/5	40	-16	64±24	29	-0.2±0.8	-0.1±0.8
	9/5-15/5	-1	0	56±21	74	-0.7±0.3	-0.8±0.3
Buskbäcken							
1980	10/4-15/4	18	0	15±15	81	0.0±0.2	0.4±0.2
	21/4-26/4	-2	0	15±15	28	-0.6±0.5	-0.5±0.5
1981	13/3- 1/4	87	87	50±50	10	3.9±5.0	4.2±5.0
	2/4-13/4	58	0	30±30	40	0.7±0.7	1.0±0.7
	14/4- 4/5	-68	-116	53±53	89	-0.0±0.6	-1.1±0.6
1982	7/4- 3/5	32	137	67±67	119	-1.5±0.6	-1.3±0.6
Gårdsjön F1							
1980	30/3- 3/4	29	0	13±13	7	2.4±1.8	2.8±1.8
	4/4- 9/4	3	0	15±15	16	-0.7±0.9	-0.7±0.9
Gårdsjön F3							
1980	31/3- 5/4	-5	0	15±15	36	-0.6±0.4	0.9±0.4
Nåsten							
1979	26/2-14/3	174	8	42±42	35	3.5±1.2	5.0±1.2
1980	15/2- 5/3	33	0	48±48	7	-2.0±6.8	-1.4±6.8
1982	23/3- 6/4	20	0	37±37	51	-0.3±0.7	-0.2±0.7
Stormyra							
1979	2/3-22/3	96	27	53±53	24	0.7±2.2	1.1±2.2
1982	2/3-25/3	78	153	60±60	86	-1.6±0.7	-1.1±0.7
	26/3- 5/4	7	-2	27±27	26	-0.7±1.0	-0.6±1.1
Svartberget							
1981	1/4- 7/4	37	0	18±18	55	0.4±0.3	0.5±0.3
	7/5-13/5	-1	0	18±18	105	-0.2±0.2	-0.2±0.2
1982	13/4-19/4	66	-28	18±18	9	8.5±1.9	8.7±1.9
	20/4-26/4	21	0	18±18	52	0.1±0.3	0.1±0.3
	27/4- 3/5	72	-38	18±18	27	3.4±0.6	3.7±0.7

$\bar{\delta}_S - \delta_M$ = difference between the $\delta^{18}\text{O}$ of snowpack and meltwater (mean values during the period).

For definitions of the other terms, see text to Eqs.(7.1-7.3).

needed for the calculations. Since M is determined from the observed changes of the water equivalent of the snowpacks, this term is uncertain for periods of little melting, particularly from large snowpacks. In such periods, the estimates of δ_M are

uncertain. Periods with very small calculated values of M have therefore been excluded. The main uncertainty in the estimates lies, however, in the calculation of A for non-melting periods and the subsequent extrapolation to melting periods. The difference given by Eq.(7.5) is $\delta_1 - \delta_M$, i.e., the difference between the initial $\delta^{18}\text{O}$ of snowpack and that of meltwater. Of more practical interest is the difference between the mean value of snowpack $\delta^{18}\text{O}$ and δ_M , listed in the last column of Table 7.4. According to these estimates, the meltwater of the investigated periods may have been either enriched or depleted as compared to the snowpack, with $\delta_s - \delta_M$ ranging between -2 and $+9\text{‰}$. These figures are weighted mean values for the periods. However, as demonstrated by Stichler et al. (1981), there may be a considerable day-to-day variation in the isotope content of meltwater, depending, among other factors, on the actual rate of melting.

Vertical ^{18}O layering of the snowpacks

As pointed out in section 6.2.6, one would expect meltwater to be depleted in ^{18}O by $1\text{--}2\text{‰}$ as compared to the snowpack. This follows from theoretical considerations regarding a homogeneous snowpack, as well as from results reported in the literature. Thus, the differences calculated above often differ from the expected ones.

One factor that largely determines the isotopic composition of the meltwater is the isotopic layering of the snowpack. The isotope and mass balances (Eqs. 7.1 and 7.2) of course also hold true for a layered snowpack. They can be used, as above, for estimates of missing terms where the other terms are measured. But for making predictions and generalized statements on the isotopic composition of meltwater, the isotopic layering has to be considered.

In this section, the observed vertical layering in some of the investigated snowpacks is presented, and the influence of the layering on the $\delta^{18}\text{O}$ of meltwater is analysed.

Examples of $\delta^{18}\text{O}$ profiles are shown in Fig.7.5. Considerable vertical differences in $\delta^{18}\text{O}$ are common in the snowpacks. The maximum difference between the sampled 5 or 10 cm layers was 9.2‰ , measured in Svartberget on March 28, 1982.

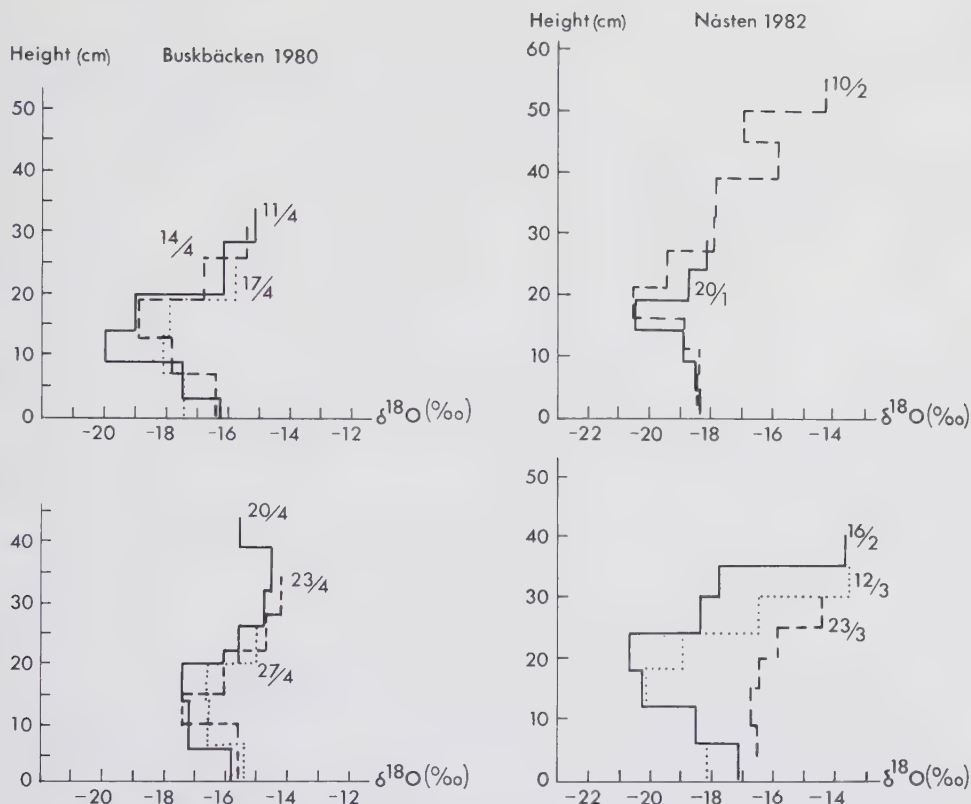


Fig.7.5 $\delta^{18}\text{O}$ -layering of snowpacks observed at different dates. Buskbäcken 1980 and Nästen 1982.

The vertical layering is a result of the variability of precipitation $\delta^{18}\text{O}$ during the snow accumulation period, as is demonstrated by Fig.7.6, showing $\delta^{18}\text{O}$ of winter precipitation (daily samples and smoothed curves) and vertical profiles in Svartberget at the end of the accumulation period of 1981/82. In order to make comparisons possible, $\delta^{18}\text{O}$ in the snowpack is plotted against accumulated water equivalent above ground. The agreements are better than was expected, in view of, among other things, the sampling difficulties for snow profiles and for snow precipitation, snow drift, and uncertain storage of the first snow of the winter. No conclusions about the rate of enrichment in various layers can be drawn, however. It is to be noted that the smooth appearance of the snowpack layering, as compared to the accumulated daily precipitation, does not indicate that the large variations in precipitation $\delta^{18}\text{O}$ have been smoothed out in the snowpack. It is merely a result of the vertical extent of the snow profile samples.

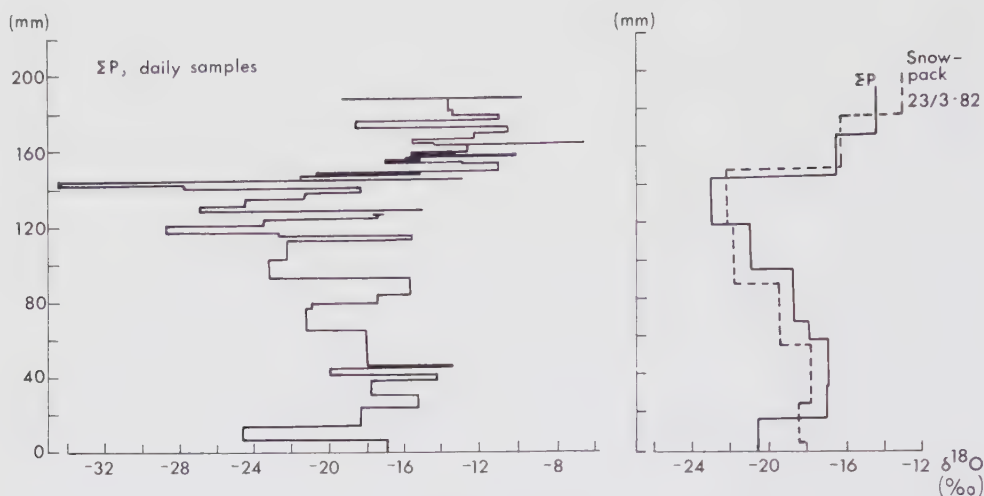


Fig.7.6 ^{18}O -layering of the snowpack compared to the $\delta^{18}\text{O}$ of winter precipitation. Vertical axes: water equivalent of precipitation or snowpack. Svartberget 1982.

A glance at Fig.7.5 shows the homogenization of the snowpack taking place during melting. It is also clear that melting takes place in the topmost layer, which is demonstrated by, for instance, the profiles from the melting period in Buskbacken 1980. The snowpack there was cut off, layer by layer, until April 18-20, when a snowfall of 19 mm occurred. This new layer disappeared during the subsequent melting.

The vertical layering of the snowpack affects the isotopic composition of meltwater to a great extent. If no fractionation processes were acting during percolation through the snowpack, and no other fractionation processes were present, the isotopic composition of meltwater would be equal to that of the melted layer (disregarding the meltwater storage in the snowpack). But the isotopic exchange during melting/freezing strives to establish isotopic equilibrium between the meltwater and each snow layer, i.e., meltwater depleted by 3‰. Probably, equilibrium is seldom reached, but the process results in a change of the original meltwater $\delta^{18}\text{O}$ and a homogenization of the snowpack.

The isotope balance of the remaining snow during a melting period of time Δt can be established according to Fig.7.7.

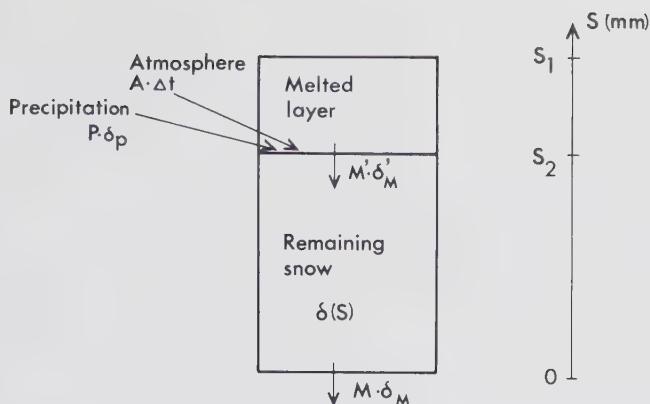


Fig.7.7 Isotope balance of a melting snowpack. For definitions of symbols, see text below.

$$\int_0^{S_2} [\delta_2(S) - \delta_1(S)] dS = M' \delta'_M + P \delta_P + A \Delta t - M \delta_M \quad (7.6)$$

- S = accumulated water equivalent from bottom to a certain level
- S_1 = total water equivalent at the beginning of the period
- S_2 = total water equivalent at the end of the period
- δ'_M = mean ^{18}O content of the layers completely melted during the period
- $\delta_1(S)$ = ^{18}O content of snow layer at level S at the beginning of the period
- $\delta_2(S)$ = ^{18}O content of snow layer at level S at the end of the period
- δ_M = mean ^{18}O content of meltwater leaving the snowpack
- $A \Delta t$ = net ^{18}O inflow from the atmosphere
- M' = volume melted in the top layer
- M = volume drained meltwater
- P = precipitation

The integral represents the storage of ^{18}O in the remaining layer. It corresponds to areas enclosed by two consecutive curves in $\delta^{18}\text{O} = f(S)$ diagrams, as exemplified by Fig.7.8 from Buskbäcken.

The balance is basically the same as the one established for mean values of snowpack $\delta^{18}\text{O}$. But in the form of Eq.(7.6), the balance makes it possible to visualize the isotopic changes during melting and percolation through the snowpack. An example is given in Fig.7.8, showing two consecutive ^{18}O profiles during melting in Buskbäcken and the various terms used in the calculations of the $\delta^{18}\text{O}$ of meltwater. The isotopically

comparatively heavy water from the melted top layer (-15.3‰) leaves some of its heavy isotopes in the snow of the deeper layers, in an attempt to establish isotopic equilibrium. Together with the flux from the atmosphere (15‰ mm) this raises the isotope content of the remaining snowpack (by 38‰ mm). According to Eq.(7.6), the process results in a depletion of the meltwater that drains from the snowpack of 0.9‰ as compared to its original value. But because of the low mean $\delta^{18}\text{O}$ of the snowpack (-17.3‰) the resulting isotope content of the meltwater (-16.4‰) is still higher than that of the snowpack.

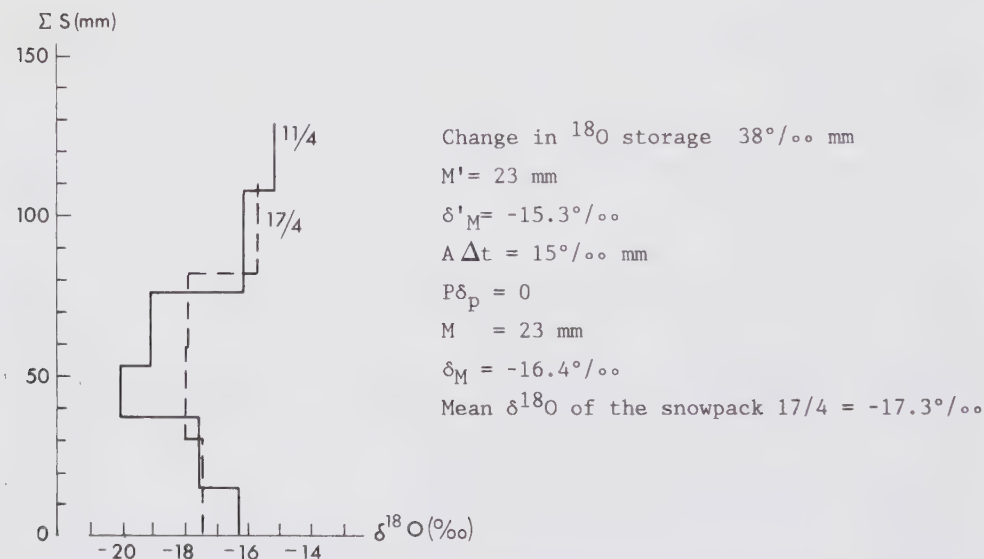


Fig.7.8 Calculation of meltwater $\delta^{18}\text{O}$ from profile changes. Buskbäcken 1980. For definitions of symbols, see p. 135.

Due to the seasonal variation in precipitation $\delta^{18}\text{O}$, the $\delta^{18}\text{O}$ of both bottom and top layers of the snowpack is probably often higher than that of the middle layers, deposited at low temperatures in midwinter (see Fig.7.6). In a deep snowpack, a high ^{18}O -content of water melting from the surface layer at the onset of melting may be reduced by isotopic exchange during percolation, and the draining meltwater may be depleted in ^{18}O as compared with the snowpack. In a more shallow snowpack, the isotopic exchange will be less, due to shorter percolation time. A high ^{18}O -content of the melting layer may then result in slightly enriched meltwater. In later meltings, of both deep and shallow snowpacks, the snowpack is isotopically more homogeneous. The ^{18}O -content of such meltwater will probably be

lower than, or similar to, the ^{18}O -content of the remaining snowpack.

7.2.2 Measured $\delta^{18}\text{O}$ of meltwater

In two of the basins investigated, Buskbäcken and Nåsten, the isotopic composition of meltwater was measured by small snow lysimeters. A 0.2 m by 0.3 m oven shelf was used in Nåsten in 1979, and 0.3 m by 0.3 m plexiglass trays were used in Nåsten and Buskbäcken in 1980. The meltwater was drained by plastic tubes and collected in plastic bottles. The lysimeters were installed below the snowpack at the onset of the melting periods, while care was taken not to disturb the snow layering. Since the snowpacks above the lysimeters were unavoidably disturbed somewhat during installation, at least by the digging and refilling of a pit on one side, these experiments are considered tentative. Due to repeated melting and freezing, combined with shallow snow depth, the experiment did not succeed in Nåsten 1980, since ice plugs were formed in the tubes.

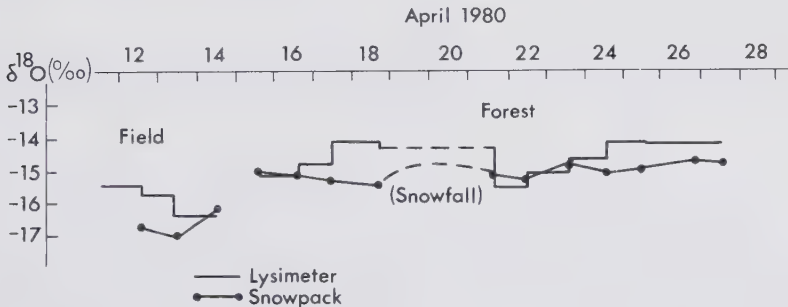


Fig.7.9 Meltwater $\delta^{18}\text{O}$ measured by lysimeters and snowpack $\delta^{18}\text{O}$ at the lysimeter sites. Buskbäcken 1980.

Fig.7.9 shows measured values of snowpack $\delta^{18}\text{O}$ close to the lysimeters and $\delta^{18}\text{O}$ of collected meltwater in the Buskbäcken experiments. Here the meltwater was usually enriched as compared to the snowpack, generally by about 0.5-1.5‰. In Nåsten 1979, meltwater was depleted as compared to the snowpack by about 1‰ during the first melting period, while the experiment failed during the second one.

In both Buskbäcken and Nåsten, the observed $\delta^{18}\text{O}$ of meltwater agrees qualitatively with the values calculated from isotope balance of single profiles close to the lysimeter sites (Table 7.5). When comparing the different results in the table, it

should be noted that the isotope balances for profiles and snowpack mean are equivalent. The two methods differ only in that the profiles (as do the lysimeters) represent single points whereas the snowpack mean represents mean values for the whole basin.

Table 7.5 $\delta^{18}\text{O}$ difference between snowpack and meltwater determined by different methods.

Basin	$\bar{\delta}_S - \delta_M$ (‰) from:			
	Lysimeter	Isotope balance of single of basin mean ¹ profile snowpack		
Buskbäcken				
1980	11/4-17/4	-0.3	-1 ± 0.7	0.4 ± 0.2
	20/4-27/4	-0.8	0 ± 0.5	-0.5 ± 0.5
Nåsten				
1979	26/2-15/3	0.5-1.3	>0	5.0 ± 1.2

$\bar{\delta}_S$ = mean $\delta^{18}\text{O}$ of snowpack during the period

δ_M = mean $\delta^{18}\text{O}$ of meltwater during the period

¹ These time periods do not fully agree with those given in the table, due to slightly different dates of samplings.

7.2.3 Snowpack $\delta^{18}\text{O}$ in forest and open sites

From the preceding discussions it follows that the isotopic composition of the snowpack within a basin should be influenced by site factors, both during the accumulation period and during the melting period. The degree of exposure to the atmosphere is an important factor governing both the net isotopic flux from the atmosphere and the rate of melting, two factors that determine the isotopic changes of the snowpack. An attempt to investigate the effect of different exposure to the atmosphere was made by comparing snowpack $\delta^{18}\text{O}$ at stations in forests with stations at open sites (fields, clear-cuttings, open bogs). In Table 7.6 the two groups are compared with respect to mean snowpack $\delta^{18}\text{O}$ at the last sampling before the melting, calculated for each basin and year, and mean $\delta^{18}\text{O}$ changes during snowmelt, calculated for the time until one of the two groups were snowfree. The snowpacks in forest sites seem to be slightly enriched, but the difference is small and not statistically significant (at the 95% level). Herrmann and Stichler

(1978), who found forest sites enriched in deuterium corresponding to a difference in $\delta^{18}\text{O}$ of about 3‰, suggested that the difference was caused by evaporation and mass exchange during the exposure of snow during interception in forests. During melting, on the other hand, the difference in Table 7.6 might indicate a higher rate of enrichment at open sites, although this difference is not statistically significant either. Higher rates of enrichment at open sites are expected due to larger turbulent exchange between snowpack and atmosphere at such sites, giving larger net isotopic flux, and due to larger net radiation causing a higher rate of melting.

Table 7.6 Comparison between the $\delta^{18}\text{O}$ of snowpack in forest and open sites.

	Forest	Open site
Mean $\delta^{18}\text{O}$ of snowpack before melting (‰)	-16.6	-17.1
standard deviation (‰)	1.8	2.0
Mean change of snowpack $\delta^{18}\text{O}$ during melting (‰)	1.9	2.4
standard deviation (‰)	1.1	2.4

The comparison is based on 11 values from each group, representing various basins and years. Each value is based on 1 to 8 observations.

7.2.4 Evaporation estimate

The net isotopic fluxes, A , calculated in Section 7.2.1, can be related to rates of evaporation by the formula for the isotopic composition of the evaporate, Eq.(6.5).

We have, for $E \neq 0$,

$$E(\delta_s - \delta_E) = A \quad (7.7)$$

where δ_E is given by Eq.(6.5).

Let us assume that the relative humidity of the atmosphere with respect to the snow surface is 0.75 and that the δ -value of the atmosphere is somewhat higher than in equilibrium with the snowpack, say $\delta_s - \delta_a = 10 \pm 5$ ‰. From Fig.6.4, redrawn with α changed to 1.015, we get $\delta_s - \delta_E \approx 40 \pm 15$ ‰. E is then related to A by

$$E = (0.025 \pm 0.011) A \text{ mm day}^{-1} \text{ for } A \text{ in } \text{‰ mm day}^{-1}$$

Thus $A = 1\text{‰ mm day}^{-1}$ corresponds to $E \approx 1 \text{ mm month}^{-1}$
 $A = 10\text{‰ mm day}^{-1}$ " " $E \approx 10 \text{ mm month}^{-1}$

The evaporation rates are seen directly from the values of A in Table 7.3, thanks to the fortuitous rough numerical agreement between ‰ mm day^{-1} and mm month^{-1} . Evaporation rates of a few millimetres per month are obtained for most periods.

These estimates may underestimate or overestimate the rate of evaporation. On the one hand, the fractionation during evaporation from snow is reduced by incomplete mixing, making the true evaporation greater than that estimated from isotopic changes by Eqs.(6.5) and (7.7). On the other hand, if the enrichment of the snowpack is caused by isotopic exchange with the atmosphere during recrystallization, as suggested in Section 6.2.6, the true evaporation may be smaller than that obtained by the equations. Due to these uncertainties and to the great sensitivity of $\delta_s - \delta_E$ to $\delta_s - \delta_a$ when h is large (Fig.6.4), the above evaporation estimates are of limited practical value, even if the rates obtained are quite realistic.

7.2.5 Conclusion on meltwater $\delta^{18}\text{O}$

It is concluded that meltwater may be either depleted or enriched in ^{18}O as compared to the snowpack at a certain moment. Due to fractionation during melting/freezing a slight depletion is expected in the general case, but vertical ^{18}O layering in the snowpack influences the difference between the δ -values. Since the areal variabilities of snowpack $\delta^{18}\text{O}$ are of a magnitude similar to the difference estimated in this section, meltwater $\delta^{18}\text{O}$ for use in the hydrograph separations in next chapter is assumed equal to that of the basin mean snowpack. Consequences of different δ -values of meltwater and snowpack are discussed in Chapter 9, using the values estimated in this section.

8 STREAMWATER. ESTIMATES OF GROUNDWATER FRACTIONS

This chapter contains the main results of the study: the estimates of the groundwater contribution to streamflow during runoff events by means of ^{18}O . A short presentation is also given of a model to estimate reservoir volumes and transit times for water in a basin by using ^{18}O .

The discussion in this chapter is mainly restricted to methodological questions arising when interpreting streamwater $\delta^{18}\text{O}$ in terms of contributions from new rain- or meltwater and groundwater. Examples of such methodological problems are the assessment of $\delta^{18}\text{O}$ of the contributing groundwater and modifications of the relationship between changes in tracer concentration and discharge during stream channel flow. Quantitative results of the separations are summarized in two tables (8.1 and 8.6) and briefly commented upon. Further conclusions concerning the process of runoff are drawn in Chapter 10.

The two-component model used, discussed in Chapter 3, reads

$$X = \frac{c_s - c_p}{c_g - c_p} \quad (3.3)$$

where X is fraction of groundwater in streamwater and c_s , c_p , and c_g are tracer concentrations in streamwater, precipitation (meltwater and rainwater) and groundwater respectively. In the applications of Eq.(3.3) for ^{18}O , the $\delta^{18}\text{O}$ of groundwater is either assumed to be constant during the runoff event or, for the snowmelt events and for certain large rainfall events, assumed to vary according to a simple mixing model (Eq.3.6). The $\delta^{18}\text{O}$ of streamwater before the runoff event is taken as the constant value of groundwater $\delta^{18}\text{O}$, or as the initial value when changes are calculated.

The occurrence of ^{18}O in precipitation and processes of isotope fractionation during water flow through a basin were discussed in Chapters 3 and 6 with the main purpose of giving background information to the use of ^{18}O for hydrograph separation during snowmelt and rainfall. The basic prerequisite for the separations is the seasonal and day-to-day variation of precipitation $\delta^{18}\text{O}$ and the fact that this variation is smoothed out in the groundwater. Thus, precipitation and groundwater $\delta^{18}\text{O}$ often

differ, making it possible to calculate the fraction of groundwater in streamwater.

From the discussions on isotope fractionation it is concluded that ^{18}O behaves like a conservative tracer during most parts of the water flow through a basin. Enrichment of importance for the hydrograph separations may, however, occur during surface water flow in certain situations (large floodings in spring or very low flow in summer in basins of little slope). Effects of this possible enrichment, on which no direct field data exist, are treated in Chapter 9. In other discussions they are disregarded.

The runoff events for which hydrograph separation has been undertaken, are marked in Figs. A1.1-9 (Appendix 1). The events are referred to by basin name, year and number within the year (snowmelt events) or letter (rainfall events). Snowmelt and rainfall events are presented separately below. The isotopic situations were comparatively similar for the different snowmelt events. The amounts of data available are also similar. The rainfall events, on the other hand, differed widely in the isotopic situation and they also differ in the amount of data available. For these reasons, the snowmelt events are treated as a group, whereas the rainfall events are presented individually. Many methodological problems, concerning both types of events, are discussed in connection with the first data presented, i.e., with the snowmelt events. Some of the snowmelt events presented here, those in Nåsten and Stormyra 1979 and those in the Gårdsjön basins 1980, were discussed by Rodhe (1980) and Rodhe (1985), respectively.

8.1 Snowmelt events

Hydrograph separation by ^{18}O was performed during 1-3 spring floods in 10 basins, making a total number of 16 separated spring floods¹. Many spring floods were composed of two or more runoff events, which were analysed separately. Thus 24 melting events were investigated.

The fraction of groundwater at the moments of streamwater sampling was calculated by Eq.(3.3). This generally gives one

¹ The term "spring flood" is used for the total streamflow caused by melting of snowpacks in spring and also for the time periods during which this flow occurs.

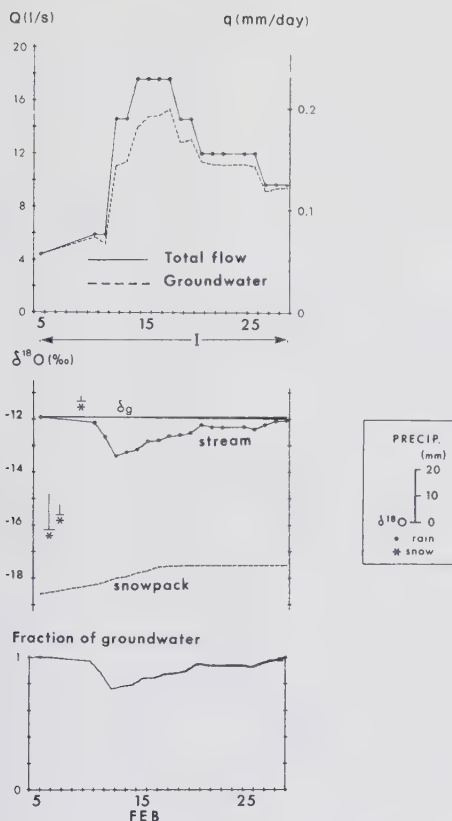


Fig.8.1 Isotopic hydrograph separation, small winter melting. Nâsten 1 1982. Stream hydrograph with calculated groundwater hydrograph, assumed $\delta^{18}O$ of groundwater (δ_g), $\delta^{18}O$ of streamwater and snowpack, amount and $\delta^{18}O$ of precipitation, and calculated fraction of groundwater.

value a day, but there are periods with more intensive as well as more sparse sampling.

The isotopic composition of meltwater was discussed in some detail in Section 7.2. It was concluded that meltwater may be either depleted or enriched as compared to the snowpack at a certain moment, depending on the vertical ^{18}O -layering of the snowpack and on the rate of melting in relation to the water equivalent of the remaining snowpack. With the data available it was not considered possible to determine the actual δ -value of meltwater accurately enough to give it a value differing from that of snowpack. Thus, meltwater $\delta^{18}O$ is assumed equal to the basin mean of snowpack $\delta^{18}O$ at any moment, and possible deviations are treated as errors. Daily values of snowpack $\delta^{18}O$ are obtained by interpolation between the days of snow sampling.

Examples of hydrograph separation by ^{18}O during snowmelt are given in Figs.8.1 and 8.2. The small runoff event in Nâsten (Fig.8.1) was caused by a short warm period, melting a few

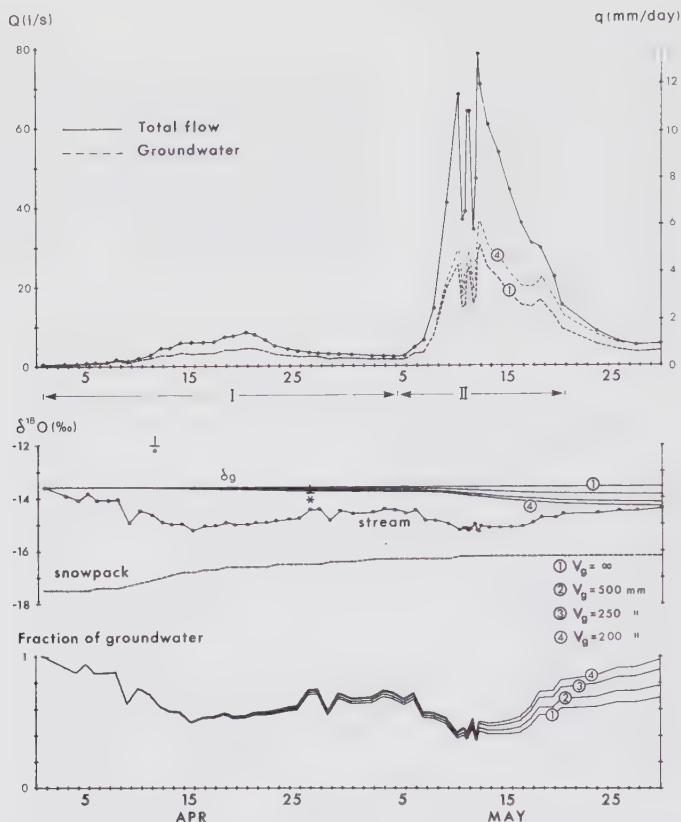


Fig.8.2 Isotopic hydrograph separation, spring flood in Svartberget Nedre 1981. The fraction of groundwater and the corresponding groundwater flow are calculated using various model reservoir volumes (V_g).

millimetres of water from a snowpack having a water equivalent of about 120 mm. The two events in Svartberget Nedre (Fig.8.2) constitute the total spring flood of the year, generated by melting of a snowpack of 200 mm water equivalent. In both examples, the error at the hydrograph separations caused by rainfall (cf. below) was insignificant, due to little rainfall during the melting, see the daily precipitation amounts and δ -values marked in the figures. During the Nåsten event, the $\delta^{18}\text{O}$ of streamwater returned to its pre-event value after the event, indicating that the $\delta^{18}\text{O}$ of the contributing groundwater did not change. In Svartberget, on the other hand, the $\delta^{18}\text{O}$ of streamwater after the event deviated considerably from the pre-event value, reflecting a change in the $\delta^{18}\text{O}$ of the contributing groundwater. Fig.8.2 shows groundwater fractions estimated using

variable $\delta^{18}\text{O}$ of groundwater, calculated by Eq.(3.6) using various reservoir volumes. Groundwater flow estimated by using the reservoir volume that yielded a change in the $\delta^{18}\text{O}$ of groundwater equal to the net change in streamwater $\delta^{18}\text{O}$ during the period is also shown (see discussion in Section 8.1.2).

Quantitative results of the hydrograph separations of the snow-melt events are summarized in Table 8.1. The table, which is frequently referred to in the following paragraphs, also gives some hydrologic data for the events. All separated spring floods are presented in Appendix 2, Figs.A2.3-19. The water equivalent of the snowpacks, the changes of which show the water input to the basins, is not shown in the "spring flood diagrams". This variable is shown in Appendix 2, Fig.A2.1, together with the observed δ -values of the snowpacks. Appendix 2 also contains a table with further quantitative results of the hydrograph separation and more detailed hydrologic characteristics of the events.

Most spring floods were composed of two, or more, distinct melting periods. The early melting was interrupted by cold weather, causing hydrograph recession until the next melting. Except for Buskbäcken in 1980, where a large snowfall occurred at the end of the main melting period, the early discharge peaks were considerably smaller than the later ones.

The different melting periods, and their related hydrographs, are referred to below by the numbers given in Figs.A2.3-19. The period boundaries are not strictly defined from the hydrographs. The boundaries rather enclose periods during which intense sampling was performed and for which hydrograph separation could be done. So are, for instance, the end of many periods chosen to minimize the influence of rainfall on the separations.

The hydrographs show discharge at the moments of streamwater sampling. Since the streams were generally sampled once a day, fluctuations in discharge within the day, often very prominent during mid and late spring flood in these streams, are not seen, except in Buskbäcken 1980 and 1981, in Gårdsjön F3 1980, and during a few intensively sampled days of the other spring floods. Momentary discharge values were chosen for the diagrams since the groundwater fraction, and thus the groundwater flow, could be calculated accurately only for the moments of streamwater sampling.

Table 8.1 Results of hydrograph separation: snowmelt events. The groundwater fractions are calculated using constant $\delta^{18}\text{O}$ of groundwater.

Basin	Period		Maximum discharge ($1 \text{ s}^{-1} \text{ km}^{-2}$)	Total runoff (mm)	Fraction of groundwater (%)	Fraction of discharge area (%)
Aspåsen	1981	I	32	51	70	12
		II	218	132	32	60
		total		183	42	37
Buskbäcken	1980	I	128	49	54	32
		II	66	32	50	33
		total		80	53	35
	1981	I	32	9	55	63
		II	123	80	57	27
		III	12	19	68	
		total		108	59	29
Gårdsjön F1	1980	total	67	42	72	38
Gårdsjön F2	1980	total	43	22	67	29
Gårdsjön F3	1980	total	71	48	80	20
Nåsten	1979	I	9	8	61	7
		II	47	54	79	18
		total		62	77	13
	1980	I	0.2	0.1		
		II	19	25		
		total		25		
	1982	I	3	3	89	2
		II	5	4	83	3
		III	108	120	55	49
		total		127	57	36

Rainfall, and also snowfall on snowfree wet areas, complicates the hydrograph separation if the precipitation $\delta^{18}\text{O}$ differs from that of snowpack. But rainfall having the same δ -value as the snowpack can be included in the term "meltwater" and snowfall, of any δ -value, on a snowpack is accounted for by the snowpack samplings. The influence of precipitation on the calculated groundwater fractions is judged for each separated event according to a procedure given in Chapter 9 and Appendix 4.

The presentation of the separated snowmelt events in the text below is organized in the following way:

Table 8.1 (Continued)

Basin			Maximum discharge ($l\ s^{-1}\ km^{-2}$)	Total runoff (mm)	Fraction of groundwater (%)	Fraction of discharge area (%)
Stormyra	1979	I	26	25	90	6
		II	76	82	85	10
		total		108	86	9
	1980	total	43	31		
	1982	total	209	148	56	49
Svartberget Övre	1981 ^a	I			40	
		II			42	
		total			41	
	1982	I	10	9	91	5
		II	98	134	57	25
		total		143	59	23
Svartberget Västra	1981 ^a	I			80	
		II			56	
		total			64	
	1982 ^a	I			95	
		II			61	
		total			64	
Svartberget Nedre	1981	I	17	21	60	14
		II	156	91	44	33
		total		111	47	28
	1982	I	7.4	15	84	13
		II	158	107	46	25
		total		122	51	25

^a The period mean groundwater fractions are calculated assuming momentary discharge values proportional to those observed at Svartberget Nedre.

First the isotopic situations of the events are commented upon.

The isotopic composition of the groundwater component, the assessment of which is an important methodological problem, is then discussed in some detail.

The next section treats methodological questions in interpreting the time variation of the $\delta^{18}O$ values of streamwater at the basin outlets. Then the variation of the calculated groundwater

fractions within the events is discussed with respect to daily values and variations within days.

Finally, the extent of discharge area estimated from the separations is presented and compared with the results of a few field surveys. The estimates of total groundwater fractions and basin fractions of discharge area are summarized briefly in this section. They are discussed and related to the flow situations in Chapter 10.

8.1.1 Isotopic situation

Due to the seasonal variation of precipitation $\delta^{18}\text{O}$, snowpacks are generally depleted in ^{18}O as compared to the streamwater. In all melting periods sampled in this study, except for those in Nåsten 1980 and Stormyra 1980, the $\delta^{18}\text{O}$ of streamwater before snowmelt differed well from that of snowpack, making possible separations by ^{18}O . Disregarding the two spring floods above, the difference at the onset of melting ranged between 2.4 and 6.2‰, with a median value of 3.9‰.

In Nåsten and Stormyra 1980 the snowpack was depleted by only 0.3‰, a situation which, together with the occurrence of precipitation of unfavourable isotopic composition, prevented hydrograph separation from being performed for these spring floods. (A very small runoff event during winter in Nåsten 1980 could be separated, Fig.A2.10. Due to uncertain relative discharge values because of very low flow and ice formation at the weir, summations were not considered meaningful, however.)

In most cases the snowpacks were continuously enriched during the periods of melting, due to drainage of isotopically light meltwater and isotopic exchange with the atmosphere and also due to precipitation (cf. Section 7.2). The difference between the $\delta^{18}\text{O}$ of streamwater before melting and that of snowpack thus generally decreased, reaching minimum values ranging between 2.1 and 4.0‰ (median value 2.7‰). As long as the snowpack $\delta^{18}\text{O}$ does not reach the assumed groundwater $\delta^{18}\text{O}$, hydrograph separation remains possible, but its sensitivity to errors in the δ -values increases as the difference diminishes (cf. the uncertainty discussion, Chapter 9).

8.1.2 Groundwater $\delta^{18}\text{O}$

As exemplified in Figs.8.1 and 8.2, streamwater $\delta^{18}\text{O}$ declined as melting started, due to intermixing with isotopically light meltwater. When melting stopped, due to cold weather or disappearing of the snowpack, streamwater $\delta^{18}\text{O}$ tended to return to its original value, indicating an increasing groundwater fraction. Before the temporal variations of streamwater $\delta^{18}\text{O}$ are analysed quantitatively in terms of the relative contribution of groundwater and meltwater, the isotopic composition of the groundwater contributing to streamflow needs some discussion, since the assessment of this quantity is a crucial point in hydrograph separation by ^{18}O .

It is regarded as impossible to calculate an accurate δ -value of the groundwater contributing to streamflow on the basis of groundwater sampling, even from a very dense network of groundwater sampling points. This is so since the relative contribution to streamflow from various reservoirs is not known. For this reason, the easily available $\delta^{18}\text{O}$ of streamwater before the runoff event is regarded as the best estimate of the required groundwater $\delta^{18}\text{O}$ at the onset of the runoff event.

As the runoff event proceeds, groundwater¹ $\delta^{18}\text{O}$ may change due to at least two processes:

1. Vertical supply by percolation of isotopically different meltwater and pre-event soil water and subsequent mixing.
2. Transformation of isotopically different soil water just above the groundwater table into groundwater by percolation and the subsequent raising of the groundwater table. Since the storage coefficient of till soil may be small (Lundin 1982), a small amount of percolating water may be sufficient to saturate a large volume of soil water and to generate a considerable rise of the groundwater table by filling a few large pores. The process results in a release of a large volume of former soil water. Due to high saturated conductivity in the near surface soil horizons, this water may constitute a considerable fraction of the total, mainly lateral, groundwater flow to the stream. The $\delta^{18}\text{O}$ of the newly formed

¹ In this particular section, the term "groundwater" refers to "groundwater contributing to streamflow".

groundwater may be dominated by the $\delta^{18}\text{O}$ of pre-event soil water in the related horizon, rather than by the $\delta^{18}\text{O}$ of the percolate.

After a sufficiently long period of streamflow recession, streamwater is probably originating from groundwater only, and streamwater $\delta^{18}\text{O}$ again equals groundwater $\delta^{18}\text{O}$. In some of the runoff events examined, streamwater $\delta^{18}\text{O}$ returned completely to its initial value at the end of the event, suggesting that groundwater $\delta^{18}\text{O}$ did not change. At the end of the main runoff events, streamwater $\delta^{18}\text{O}$ often deviated considerably from its initial value, indicating a change of groundwater $\delta^{18}\text{O}$.¹

Unfortunately, for the sake of isotopic hydrograph separation, the temporal variation of groundwater $\delta^{18}\text{O}$ within the event is not known, only its initial and final values. Any estimate of the variation implies a concept of the process of groundwater recharge and flow to the stream. Even if streamwater $\delta^{18}\text{O}$ after the event equals the pre-event value, different interpretations can be made:

If mixing with percolating water (case 1 above) or piston flow dominates, the return of streamwater $\delta^{18}\text{O}$ to its original value indicates that groundwater $\delta^{18}\text{O}$ has remained constant during the event, due to sufficiently large reservoirs being involved.

If, on the other hand, activation of isotopically different soil water dominates (case 2 above), groundwater $\delta^{18}\text{O}$ may have been changing within the event, even if the final value equals the initial one. In this case the isotopic composition of the discharging groundwater is a function of the height of the groundwater level as long as the pre-event soil water dominates the groundwater flow in the upper layers. When the groundwater level falls after the event, the groundwater $\delta^{18}\text{O}$ returns to its initial value.

Assuming that case 1 applies, and further that the $\delta^{18}\text{O}$ of pre-event soil moisture equals that of the groundwater, the use of constant groundwater $\delta^{18}\text{O}$ in separation gives a lower limit for the groundwater fraction. What is actually calculated is the fraction of pre-event groundwater or soil water. If streamwater after the event equals the initial value, the true groundwater

¹ Precipitation on saturated areas is probably responsible for the rise of streamwater $\delta^{18}\text{O}$ at the end of some of the events. These cases are excluded from the quantitative comparisons between the basins below.

fraction is obtained. The larger the difference between initial and final streamwater $\delta^{18}\text{O}$, the larger will be the underestimation of the groundwater fraction when using constant groundwater $\delta^{18}\text{O}$.

In case 2, the groundwater $\delta^{18}\text{O}$ may change in either direction within the event, even if the final value equals the initial one. The use of constant $\delta^{18}\text{O}$ of groundwater may then cause overestimation as well as underestimation of the groundwater fraction. Overestimation, i.e., the occurrence of higher groundwater $\delta^{18}\text{O}$ within the event, is unlikely during spring floods, however. This is so since autumn rainfall, which to a large degree determines the isotopic composition of soil water during winter, is unlikely to have higher $\delta^{18}\text{O}$ than the groundwater.

A close look at the time courses of streamwater $\delta^{18}\text{O}$ (Figs. A2.3-19) reveals some characteristic dissimilarities among the events and among the basins, as to changes of groundwater $\delta^{18}\text{O}$. The spring floods seem to give changes of groundwater $\delta^{18}\text{O}$ of 1-2‰. The first small runoff events exert various influence on groundwater $\delta^{18}\text{O}$. In Nåsten and Stormyra 1979, and in Svartberget Västra 1981, small or no changes are encountered, while changes seem to occur in Aspåsen, Buskbäcken, and Svartberget Övre. (Svartberget Nedre, being largely a mixture of Svartberget Övre and Svartberget Västra, is not treated separately here.)

Changes in groundwater $\delta^{18}\text{O}$ within an event depend on isotopic and flow conditions as well as on basin physiography. If groundwater $\delta^{18}\text{O}$ is changed mainly by mixing with infiltrating water, the total change should increase with the difference between groundwater and meltwater $\delta^{18}\text{O}$ and with meltwater amount. For a completely mixed reservoir of constant volume, V_g , the rate of change is given by Eq.(3.6)

$$\frac{d\delta_g}{dt} = \frac{Q}{V_g} (\delta_p - \delta_g) \quad (3.6)$$

For the hydrograph separations, the reservoir volumes in Eq.(3.6) can be chosen by trial and error to simulate the total change in groundwater $\delta^{18}\text{O}$. From Figs.A2.3-19 it is seen that model reservoir volumes of 100-300 mm often give the desired change in groundwater $\delta^{18}\text{O}$ during the spring flood period. The time courses of the calculated groundwater $\delta^{18}\text{O}$, on the other hand, show that except for in Aspåsen 1981 and Stormyra 1979 smaller reservoir volumes are needed to simulate the changes during the initial runoff events. Simulations of groundwater

$\delta^{18}\text{O}$ with various values of V_g give reservoir volumes for many of the initial periods of around 40-100 mm.

Regarding the above results on model reservoir volumes and further the basic relationship between groundwater level and flow, which should cause vertical as well as lateral expansion of the contributing groundwater reservoir as the flow increases, a flow-dependent reservoir volume is probably more realistic than the constant volume applied here. But in a desire to minimize the number of assumptions and non-measurable parameters in the interpretation of streamwater $\delta^{18}\text{O}$, the present analysis is restricted to constant model reservoir volumes.

Observed $\delta^{18}\text{O}$ of groundwater

To give an idea of the assumptions concerning the $\delta^{18}\text{O}$ of the groundwater component, a few data on the areal and temporal variations of groundwater $\delta^{18}\text{O}$ in the basins are presented below.

An example of the areal variability of groundwater $\delta^{18}\text{O}$ is given in Table 8.2, showing the results of two synoptical surveys in the Gårdsjön basins. The groundwater was sampled in shallow observation wells in the loose deposits, mainly in discharge areas. The surveys were performed at summer low and autumn medium groundwater stages. The standard deviations, 0.2-0.4‰, may be indicative of the accuracy to be expected of the $\delta^{18}\text{O}$ of groundwater contributing to streamflow.

Table 8.2 Synoptical survey of groundwater $\delta^{18}\text{O}$ within and close to Gårdsjön F1, F2, and F3

Date	Number of observations	Groundwater $\delta^{18}\text{O}$		
		mean	stand. dev. ($^{\circ}/_{\infty}$)	$\delta_{\max}-\delta_{\min}$
18/7 1982	22	-10.27	0.19	0.5
5/11 1982	20	-9.00	0.44	1.7

Examples of changes over time of groundwater $\delta^{18}\text{O}$ during runoff events are shown in Fig.8.3 (see also examples from the snowmelt events in the Gårdsjön basins in Rodhe (1985)). Comparatively large changes are seen in these springs and wells. In all these events, however, groundwater $\delta^{18}\text{O}$ was changing in the direction expected from the isotopic composition of the infiltrating

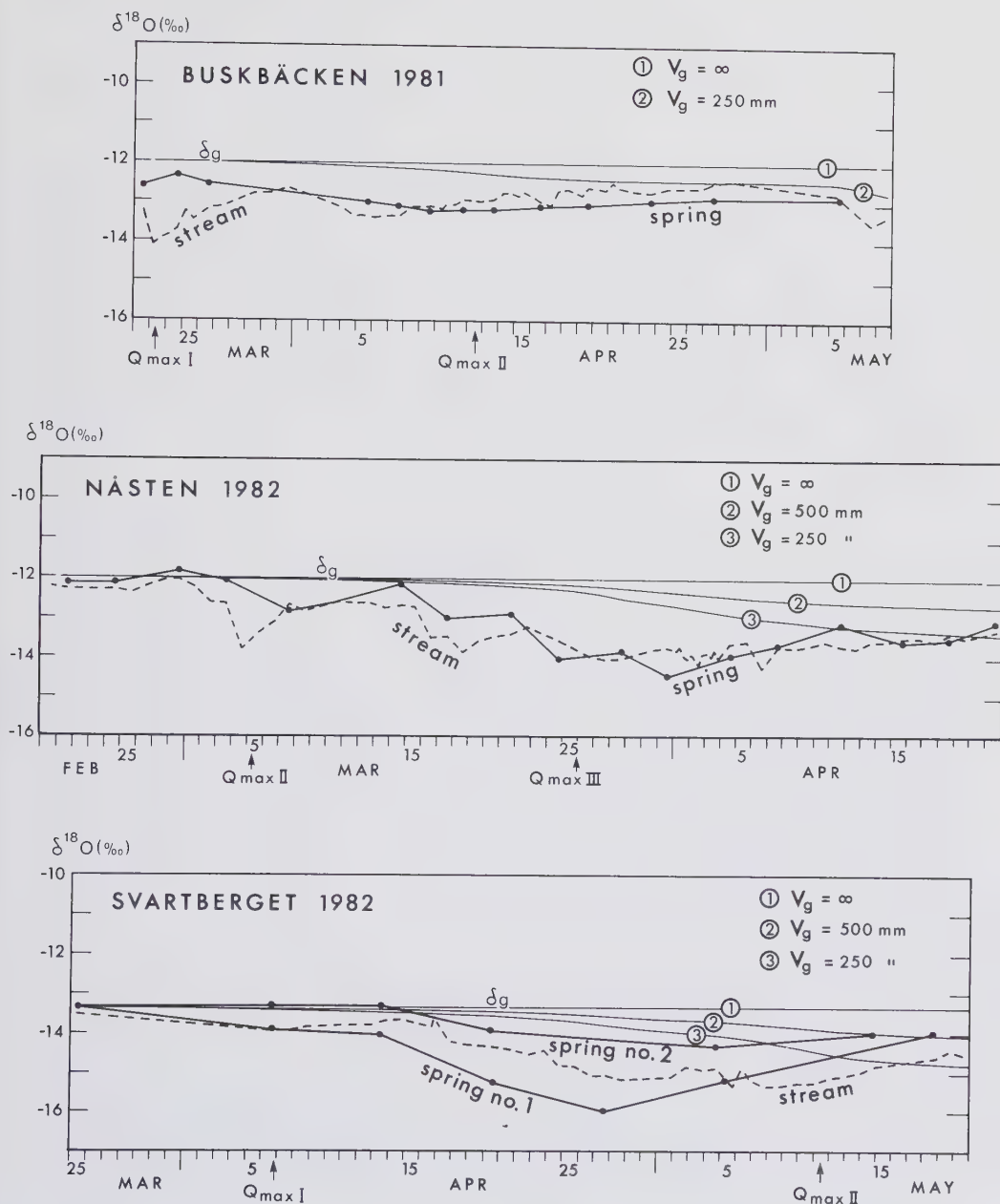


Fig.8.3 $\delta^{18}\text{O}$ of groundwater in springs, calculated $\delta^{18}\text{O}$ of the groundwater component of streamflow (δ_g) for various model reservoir volumes (V_g), and $\delta^{18}\text{O}$ of streamwater. Snowmelt events. The dates for event maximum streamflow are marked by Q_{max} .

water. This fact supports the above statement that the use of the initial streamwater $\delta^{18}\text{O}$ as constant groundwater $\delta^{18}\text{O}$ yields lower limits of groundwater fractions in streamwater.

The close agreement between the δ -values of groundwater and streamwater in some of the cases is surprising. Some of the springs and wells even show a larger influence from precipitation than does the streamwater. Obviously there must be some other groundwater that contributes to streamflow, since streamwater $\delta^{18}\text{O}$ must lie somewhere in between the $\delta^{18}\text{O}$ of groundwater and of snowpack. Considerably smaller changes over time were found in samples from deeper wells and borings.

Application of Eq.(3.3) on the data in Fig.8.3 gives fractions of newly infiltrated meltwater of up to 80% in the Buskbäcken spring, 60% in the Nåsten spring, and 60% in the Svartberget springs.

The rise in spring water $\delta^{18}\text{O}$ occurring in the later parts of the periods is noticeable. It shows that neither piston flow nor case 1 above, i.e. one well-mixed reservoir, describes their isotopic behaviour. It may reflect the fact that water infiltrating close to the spring, where the unsaturated zone is thin, constitutes a large fraction of the groundwater flow to the spring as long as infiltration is going on. When the infiltration ceases, the discharge becomes dominated by older groundwater and groundwater recharged during the event at more distant localities, where the unsaturated zone is thicker and where the recharging soil water particles were present in the soil already before the event.

Conclusions on groundwater $\delta^{18}\text{O}$

From the above discussion on $\delta^{18}\text{O}$ of groundwater contributing to streamflow it is concluded that:

1. Hydrograph separation using the initial streamwater $\delta^{18}\text{O}$ as constant groundwater $\delta^{18}\text{O}$ during the event yields lower limits of groundwater fractions in streamflow. The degree of underestimation can be judged from the total changes in streamwater $\delta^{18}\text{O}$ during the event.
2. The assumption of complete mixing of the infiltrated meltwater, in a reservoir of constant volume, gives too small rate of change in calculated groundwater $\delta^{18}\text{O}$ at the

beginning of the period. Hydrograph separation using calculated groundwater $\delta^{18}\text{O}$, with a model reservoir volume chosen to simulate total change in groundwater $\delta^{18}\text{O}$ during the event, thus also underestimates the groundwater fraction at runoff events where groundwater $\delta^{18}\text{O}$ changes.

8.1.3 Modification of changes in tracer concentration and discharge during flow in the channel network

The estimated groundwater fractions of the total water volumes discharged in the streams during the various snowmelt events, i.e., the period mean fractions weighted with respect to discharge, were given in Table 8.1. In a following paragraph, the temporal variation of the estimated groundwater fractions within the events is discussed. In order to distinguish the two aspects of the groundwater fractions, the term "instantaneous groundwater fraction" or just "groundwater fraction" will be used for fractions calculated from single observations of streamwater $\delta^{18}\text{O}$, whereas the event mean values will be referred to as "total groundwater fractions".

Before analysing the temporal variations of the instantaneous groundwater fractions, modifications of the relationship between the changes in tracer concentration and discharge that may occur during flow in the stream channel network are discussed.

The groundwater fractions refer to fractions calculated from the isotopic composition of streamwater at the basin outlets. When the groundwater fractions are to be interpreted in terms of groundwater and meltwater inflow to the stream along its reaches, and variations of these inflows over time, the modifying effects of the channel network upon changes in tracer concentration and discharge have to be considered.

Firstly, the difference between the propagation rate of flow disturbances and of changes in tracer concentration, i.e., between flood wave celerity and water particle velocity, has to be considered for an analysis of the relationship between groundwater fraction and total flow within the events.

Secondly, the damping of input fluctuations in tracer concentration during channel flow and flow through surface water reservoirs has to be considered for an analysis of diurnal fluctuations in groundwater fraction.

The above two effects are briefly treated in this section as a background to the analyses of the variations of streamwater $\delta^{18}\text{O}$ and calculated groundwater fractions within the events.

Relationship between flood wave celerity and particle velocity

The celerity of a kinematic flood wave in a channel, c , is related to the velocity of the water particles, v , by

$$c = v + A \frac{dv}{dA} \quad (8.1)$$

where A is the cross-sectional area of the channel (Henderson 1966, p.366). By the de Chézy or Manning formula, v can be related to channel geometry giving

$$c = \text{const} \cdot v \quad (8.2)$$

with the constant having values around 1.5. It follows from the equation that a change in water level and discharge is propagated downstream faster than a change in tracer concentration. As pointed out by Glover and Johnson (1974), this velocity difference, if not accounted for, may lead to erroneous conclusions when interpreting the relationship between the concentration of chemical compounds and stream discharge in terms of runoff processes in the basin.

Eq.(8.2) has been verified in several studies, e.g., Glover and Johnson (1984), but mostly in stream channels being larger and more regularly shaped than the small forest streams of the present study. Therefore a few experiments were performed in two of the present study basins, Aspåsen and Buskbäcken, and also in a similar basin, using an artificial tracer. The flow changes were generated by pumping of water to or from the streams. In the 3 experiments, the ratio c/v ranged from 1.5 to 2.5. These results differ from those of Pilgrim (1977) who found c/v around 1.0 in small Australian forest streams.

Eq.(8.2) presumes that the tracer is transported with the mean velocity of the water in the cross-section of the stream. This may not be the case during bank overflow when the velocity is much lower in the flooded areas than in the stream channel itself. In such a situation Walling and Webb (1980) observed values of $c/v < 1$, i.e., tracer changes propagating more rapidly than flow changes.

With $c/v = 1.5$ we get the upper limit of the time lag between flow and tracer changes, $t_{1,max}$, in a basin as

$$t_{1,max} = t_f - t_c \leq 0.33 t_f \quad (8.3)$$

where t_f is maximum floating time and t_c time of concentration of the channel net, i.e., the maximum time required for a flow disturbance in the channel net to reach the basin outlet.

The velocity of the water particles was measured using artificial tracers in Aspåsen and Buskbäcken. The velocity increased downstream from around $0.05 \text{ m}\cdot\text{s}^{-1}$ at the upper part of first order streams to around $0.2 \text{ m}\cdot\text{s}^{-1}$ at the well-developed stream channel at the outlet of the Buskbäcken basin. If these velocities, observed during periods of medium flow, are assumed representative for all basins in the study, floating times ranging from less than 0.02 days in the Gårdsjön basin to about 0.5 days in Nåsten and Stormyra are obtained. The corresponding maximum time lags are, by Eq.(8.3), less than 0.2 days in all basins. Regarding the actual continuous inflow along the stream, rather than the case of point inflow at the most remote part of the channel net treated so far, it follows that the actual time lag is probably considerably less than 0.2 days.

It is concluded that the time lag between flow change and tracer change can be disregarded at the outlets of all basins when daily values of flow and tracer concentration are compared. Caution may be needed, however, when hourly values are compared in the largest basins.

Damping and delay of fluctuations in tracer concentration

The tracer concentration, as well as the flow, at a certain moment at the sampling point is the integrated result of inflows along the stream that have taken place at different times. This fact, together with the storing of water and tracer in the channel and occasional surface water reservoirs (flooded areas, ponds, etc.), damps and delays input fluctuations in flow and tracer concentrations.

The existence of large diel¹ fluctuations in discharge during snowmelt in the investigated basins, which is commented on in the next section, shows that diel fluctuations in the inflow along the streams are propagated to the basin outlets. With

¹ diel = within the day (24 hours)

time lags of the tracer of at most 0.2 days, i.e., 0.2 period lengths, it follows immediately that diel fluctuations in tracer concentration are also propagated along the stream. There is a delay and damping of the input fluctuations of both flow and tracer, but the degree of damping cannot be estimated without a closer analysis, which has not been performed here.

The above discussion concerns the effect of stream channel flow. During flow through surface reservoirs there may be a considerable damping of diel fluctuations in tracer concentrations. This is shown by simulations of the isotopic composition of the outflow from the completely mixed reservoirs characterized in Table 8.3, assuming sinus-shaped fluctuations in flow and tracer concentrations of the inflow. The isotopic composition of the outflow is given by Eq.(6.12) with the inflow (I), the $\delta^{18}\text{O}$ of the inflow (δ_I) and the reservoir volume (S_0) varying, and $E = 0$. The surface area of the reservoirs is held constant, and the outflow is calculated from reasonable or actual rating curves.

In Stormyra the amplitude in tracer concentration in the output was 25% of the input value, while in Gårdsjön F3 it was 5-10% of the input value. Since the mixing in reality is not complete, the actual damping is less.

Table 8.3 Flow situations for which damping in reservoirs is estimated.

	Mean discharge (l s^{-1})	Reservoir area (ha)	Initial depth (m)
Stormyra	200	5	0.1
Gårdsjön F3	2	0.2	0.1

It is concluded that occasional surface reservoirs may very significantly reduce diel fluctuations in the isotopic composition of streamflow. Surface reservoirs capable of damping the fluctuations in streamwater concentration are found during the spring flood in Gårdsjön F3, Stormyra, and to some extent in Nåsten, but not in the other basins.

8.1.4 Variations of groundwater fractions within the events

The hydrograph separations by ^{18}O show that groundwater accounts for a considerable part of total streamflow during the spring floods. As a whole there is an inverse relationship between groundwater fraction and stream discharge. Minimum calculated instantaneous groundwater fractions for all spring floods studied range between 0.17 and 0.73 with a median value around 0.4. Except in Buskbäcken 1981 and in Nåsten 1979 the total minima occurred around the date of spring flood culmination.

Daily variations in groundwater fractions in streamflow

The roughness of the assumption made in this study, that melt-water $\delta^{18}\text{O}$ is equal to snowpack $\delta^{18}\text{O}$ at a certain moment, prevents a very detailed analysis of the time courses within the events of the calculated groundwater fraction. It is still considered that some general features can be pointed out which may elucidate the process of runoff. Three common behaviours of the groundwater fractions can be identified.

Firstly, an inverse relationship between groundwater fraction and total flow is found within all runoff events, as is seen in the spring flood diagrams and more explicitly in the examples shown in Fig.8.4.

Secondly, most plots of groundwater fraction versus total flow for the whole spring flood period show a change of slope of the curves at a certain, low, discharge (see Fig.8.4). During the initial peaks, and sometimes during the early rise of the main peaks, the curves have a more negative slope than at higher discharges. For some of the spring floods, the slope in the diagram is zero at high discharges, i.e., the groundwater fraction is independent of total flow.

Thirdly, most initial melting periods show a time lag between groundwater fraction changes and flow changes. A close look at these periods in the spring flood diagrams shows that the calculated minimum groundwater fractions often occur a few days before the discharge peak, see for instance Aspåsen I 1981 and Nåsten I 1982. The time lag appears as a counter-clockwise hysteresis effect in the plots of groundwater fraction versus total flow (Fig.8.4).

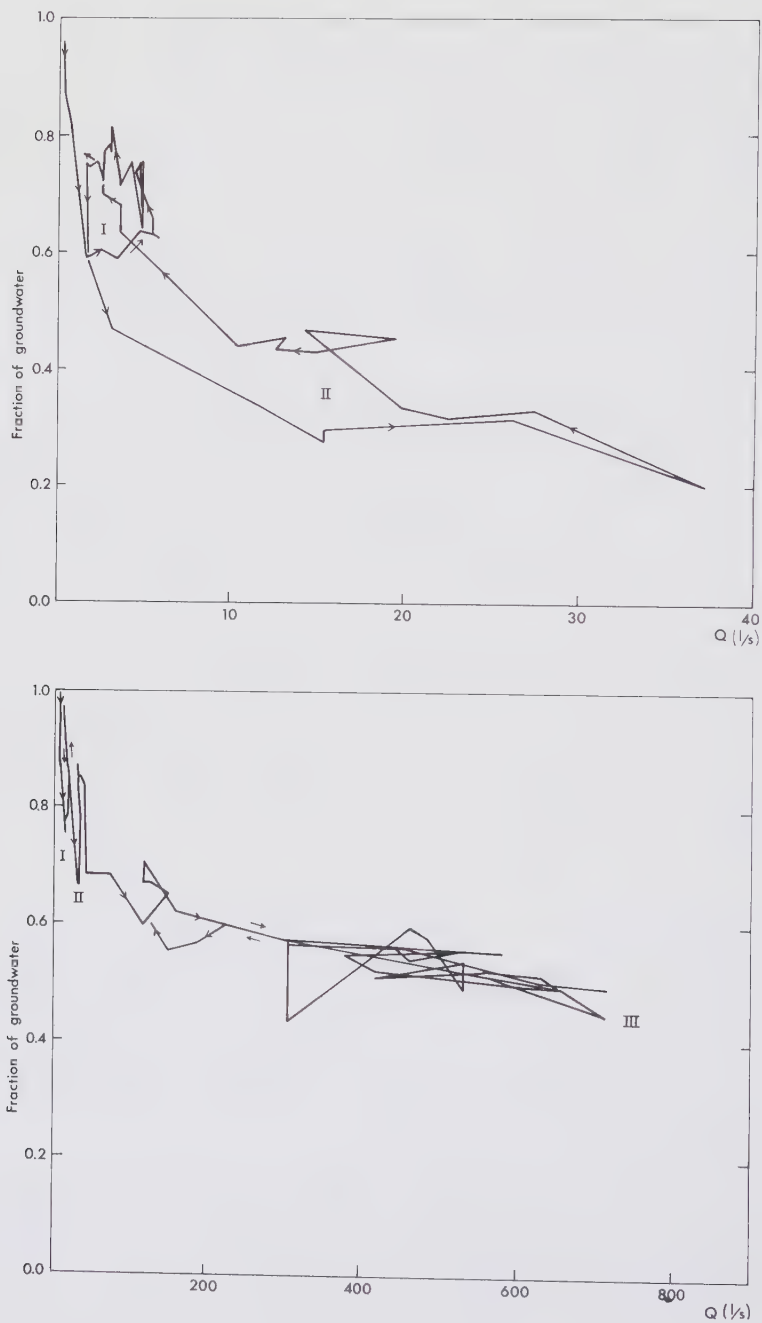


Fig.8.4 Fraction of groundwater (constant δ_g) versus stream-flow. Spring floods in Aspåsen 1981 (upper diagram) and Nåsten 1982 (lower diagram). Roman numerals indicate snowmelt event number (see Figs.A2.3 and A2.12).

Also some main melting periods show a time lag between groundwater fraction and total flow, see Fig.8.4 for an example. These lags are less pronounced, and the peaks are contemporary in most cases. The more realistic groundwater fractions, calculated by using variable groundwater $\delta^{18}\text{O}$, yield a still more pronounced hysteresis effect, Fig.8.5. Small opposite time lags, i.e., changes in groundwater fractions being delayed as compared to those of total flow, were found in some main periods (e.g. Stormyra 1982). But when variable groundwater $\delta^{18}\text{O}$ was used, changes in groundwater fractions preceded flow changes in most of these cases.

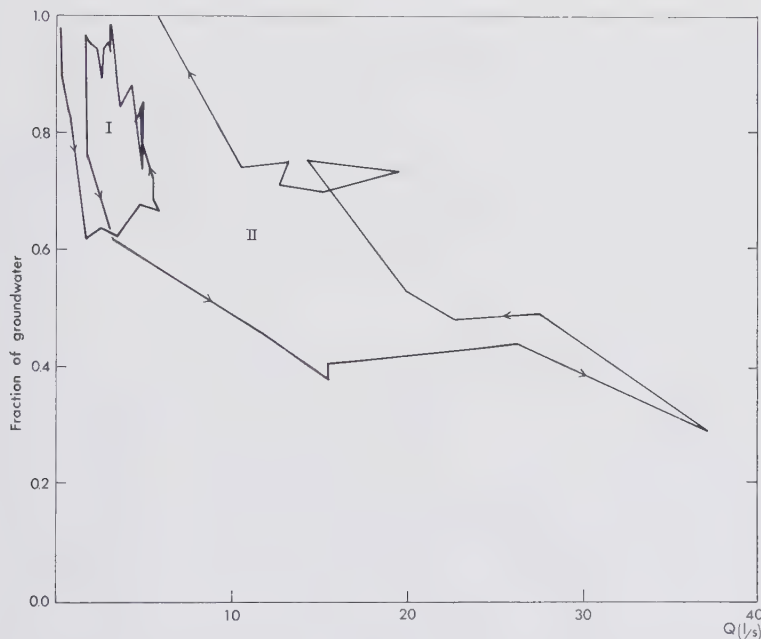


Fig.8.5 Fraction of groundwater (variable δ_g , $V_g = 250$ mm) versus streamflow. Spring flood in Aspåsen 1981.

From the discussion on different transport velocities of changes in tracer concentration and discharge in Section 8.1.3, it was concluded that if there is any difference between these velocities, it is probably a delay of the tracer change, this delay being less than 0.2 days. The observed time lag between groundwater fraction and total flow at the basin outlets, showing that tracer change precedes flow change, thus reflects a real time lag during water inflow to the stream. The time lag implies that meltwater that reaches the stream directly does so slightly more rapidly than the bulk of the groundwater contributing to the peaks.

The rapid decline in calculated groundwater fraction at the beginning of the initial peaks may partly be fictitious, due to a methodological error. Since the most depleted meltwater from an unlayered snowpack is to be expected at small rates of melting from a deep snowpack, the actual meltwater $\delta^{18}\text{O}$ at the onset of melting might be a few per mil lower than the snowpack mean. As melting proceeds, this difference is likely to decrease. But the general tendencies of the time variations in the groundwater fractions pointed out above are unlikely to have been caused only by errors in the assessment of meltwater $\delta^{18}\text{O}$.

Diel fluctuations in groundwater fractions

The basins investigated are all small enough to show considerable diel fluctuations in discharge during snowmelt. At the dates of maximum flow, the relative amplitude¹ in discharge during the day, defined as $(Q_{\text{max}} - Q_{\text{min}})/(Q_{\text{max}} + Q_{\text{min}})$, often was around 0.25. The largest value at maximum flow, 0.39, was found in Svartberget Nedre 1981 and the smallest, 0.12, in Stormyra 1982.

When diel fluctuations in streamwater $\delta^{18}\text{O}$ are to be interpreted, two considerations have to be made. Firstly, the damping effect of the channel net and occasional surface reservoirs upon fluctuations in input concentrations has to be regarded. As discussed above, damping may reduce input fluctuations in the largest basins and in the basins where considerable surface-water reservoirs develop, i.e., in Gårdsjön F3, Nåsten, Stormyra and Svartberget Övre.

Secondly, the groundwater fractions of the daily contributions to streamflow must be distinguished from those of the total streamflow, i.e., from those shown in the spring flood diagrams. Before discussing the intensively sampled events, a relationship between these two groundwater fractions is deduced.

Consider the stream hydrograph and the groundwater hydrograph, as estimated by ^{18}O , in Buskbäcken during the days around the culmination of the spring flood in 1980 (Fig.8.6). According to the hydrograph separation by ^{18}O the groundwater fraction at

¹ The amplitude is the maximum deviation from the mean value. The total variation is thus twice the amplitude. Q_{max} and Q_{min} refer to maximum and minimum discharge values during the day.

the hour of maximum flow on April 14 was 0.44, giving a groundwater flow of 103 l/s. A considerable part of this groundwater flow, however, was generated on the preceding days. Assuming a groundwater recession curve as sketched in the figure, the groundwater fraction of the flow generated on April 14, X' , becomes

$$X' = \frac{Q X - Q_{g,r}}{Q - Q_{g,r}} = \frac{X - \frac{Q_{g,r}}{Q}}{1 - \frac{Q_{g,r}}{Q}} \quad (8.4)$$

where Q is the total flow, $Q_{g,r}$ is the groundwater recession flow, and X is the groundwater fraction of total flow. The calculated values of X' are shown in the figure. According to these calculations, the groundwater fraction of the daily contribution was 0.33 at the hour of maximum flow, which value should be compared with the groundwater fraction of total flow, 0.47.

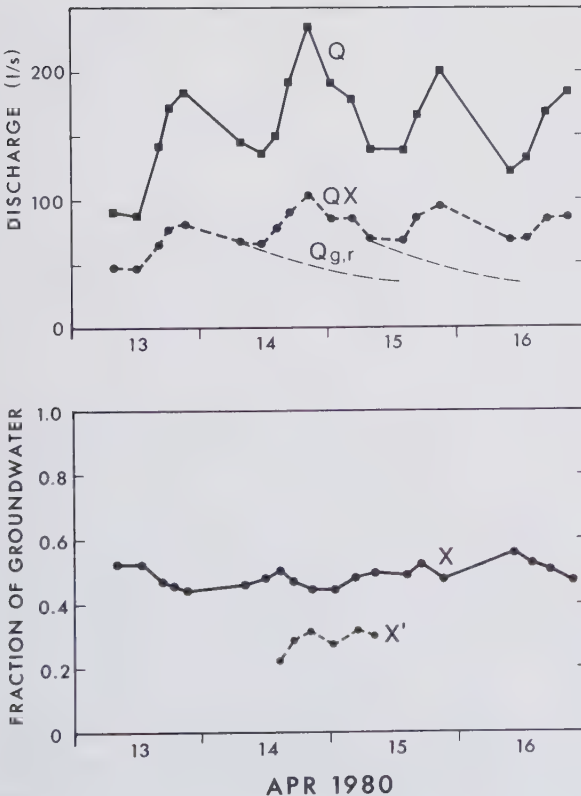


Fig.8.6 Identification of the groundwater fraction of the daily contribution to streamflow.

Q = total streamflow,
 QX = groundwater flow,
 $Q_{g,r}$ = groundwater recession flow from the preceding day,
 X = groundwater fraction of total streamflow,
 X' = groundwater fraction of the daily contribution to streamflow.

Data from Buskbäcken 1980.

This distinction between the two groundwater fractions may also be of importance when looking at a whole event. In the discussion above it was pointed out that there is an inverse relationship between groundwater fraction and total flow within the snowmelt events. By definition, however, the groundwater fraction is 1.0 at the onset of the events. Any change in the groundwater fraction must accordingly start with a reduction of this fraction. A more stringent treatment would have been to analyse the groundwater fraction of the flow increment during the events as calculated by Eq.(8.4). But when whole melting periods are considered, the pre-event flow is very small compared to the flow during the events. Thus $Q_{g,r}/Q$ is close to zero and X' is approximately equal to X . This is true also for the majority of the rainfall-generated events presented in Section 8.2 and Appendix 3. Noting the distinction between the two groundwater fractions, the following discussion on diel fluctuations in groundwater fractions during snowmelt events is restricted to the groundwater fractions of the total streamflow, X .

Most spring floods were sampled only once a day, but Buskbäcken 1980 (of which a few days were shown in Fig.8.6) and Gårdsjön F3 1980 were sampled about four times a day, and Buskbäcken 1981 twice a day. All spring floods studied in 1981 and 1982 were intensively sampled during a few days near the day of maximum flow.

Fig.8.7 compares total discharge and $\delta^{18}O$ of streamwater during days with fluctuating discharge in Buskbäcken 1980 and 1981. To simplify the comparisons, the positive $\delta^{18}O$ -axes are directed downward in the diagrams. In 1980 there was a small daily oscillation in streamwater $\delta^{18}O$ being inverse to that of discharge, indicating the largest relative contribution of meltwater at the hour of maximum discharge. The amplitudes in $\delta^{18}O$ were about 0.1-0.2‰, giving amplitudes in calculated groundwater fraction of total flow of about 0.02-0.05. In Buskbäcken 1981, on the other hand, no significant difference in $\delta^{18}O$ and calculated groundwater fraction was found between the two hours of sampling, largely representing maximum and minimum daily flow. The finding is surprising since the relative (and absolute) amplitudes in discharge were largest in 1981. A corresponding dissimilarity between the two years was also found in the electrical conductivity of the streamwater. The different behaviour could not be related to differences in frost depth or soil moisture upon freezing. Since the total discharge was about the same for the two years, different damping in the channel

network is unlikely to explain the different isotopic behaviour. The difference thus reflects a difference in the diel variations of the groundwater and meltwater flow to the stream. It should be noted, however, that also in 1980 the relative variations of the groundwater fractions were much smaller than those in total discharge, yielding large diel fluctuations in the calculated groundwater flow both years (see Figs.A2.4 and A2.5).

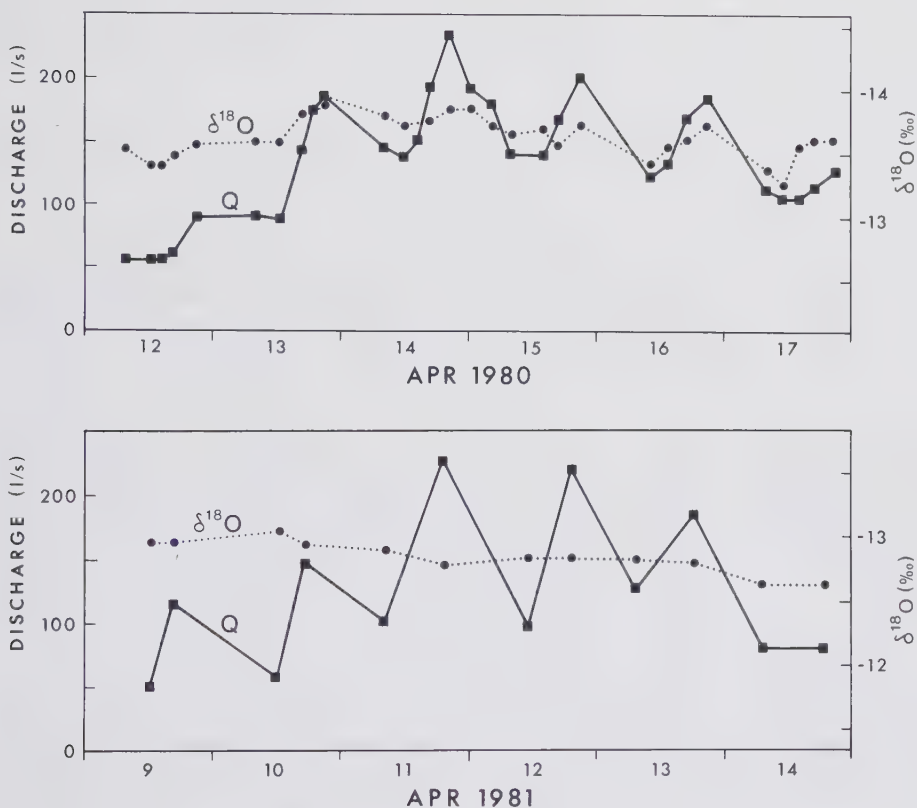


Fig.8.7 Comparison between diel variations in discharge and $\delta^{18}\text{O}$ of streamwater. Spring floods in Buskbäcken 1980 and 1981. (N.b. the downward direction of the $\delta^{18}\text{O}$ -axes.)

In Gårdsjön F3 1980 (Fig.A2.8), which had similar relative fluctuations in discharge as Buskbäcken 1980, no significant fluctuation in streamwater $\delta^{18}\text{O}$ was noted. This indicates amplitudes in the groundwater fraction of total flow of less than 0.02.

Results from all intensively sampled periods are summarized in Table 8.4. Since the intensive sampling was usually not undertaken exactly on the day of spring flood culmination, the daily

mean and relative amplitude of discharge are given both for the day of maximum flow and for the day of intensive sampling.

As is seen in Table 8.4, the diel variations in calculated groundwater fractions of the total flow are comparatively small, around 0.05-0.10. In basins where a measurable diel variation in streamwater $\delta^{18}\text{O}$ was found, the calculated groundwater fractions were found to be inversely related to total flow, as expected. Diel fluctuations in $\delta^{18}\text{O}$ were usually accompanied by similar fluctuations in electrical conductivity, showing the same tendency as $\delta^{18}\text{O}$.

Table 8.4 Diel fluctuations in discharge, $\delta^{18}\text{O}$ of streamwater, and groundwater fractions during snowmelt. Observations during the days of intensive sampling.

Basin		Daily mean specific discharge ($\text{l s}^{-1} \text{ km}^{-2}$)	Relative amplitude in discharge	Amplitude in $\delta^{18}\text{O}$ ($^{\circ}/\text{‰}$)	Daily mean groundwater fraction, X	Amplitude in X, ΔX
Aspåsen	1981	100 (204) ^a	0.15 (0.24)	0.08	0.31	0.01
Buskbäcken	1980	103	0.25	0.23	0.50	0.03
	1981	90	0.40	0.06 ^b	0.57	0.01 ^b
Gårdsjön F3	1980	57	0.22	0.19 ^b	0.80	0.03 ^b
Nåsten	1982	83	0.11 (0.24)	0.28	0.50	0.03
Stormyra	1982	83 (194)	0.08 (0.12)	0.41	0.53	0.06
Svartberget Övre	1981			0.18	0.36	0.03
	1982	53 (≈ 100)	0.21	0.11	0.40	0.02
Svartberget Västra	1982			0.19	0.57	0.02
Svartberget Nedre	1981	104	0.39	0.27	0.41	0.06
	1982	61 (120)	0.23	0.38	0.56	0.05

^a Values within brackets refer to the situation at maximum flow

^b No periodic fluctuation, rather accuracy of measurement

The absence of large fluctuations in streamwater $\delta^{18}\text{O}$ in Svartberget Nedre 1981 is surprising. This spring flood showed very regular fluctuations in discharge. The intensive sampling was performed on the days of maximum flow, having a relative

amplitude in discharge of 0.39, see Fig.8.2. Damping of any significance in the stream channels is unlikely in these streams. The lack of fluctuations in $\delta^{18}\text{O}$ at Svartberget Övre might be explained by the large fraction of mire in this sub-basin. But the flow from Svartberget Övre is only about one third of the total flow at Svartberget Nedre. Thus the stream is fed by groundwater and meltwater in roughly constant proportions during the day.

Conclusions on diel fluctuations

One important methodological conclusion can immediately be drawn. The absence of large diel fluctuations in streamwater $\delta^{18}\text{O}$ shows that daily sampling is often sufficient during snow-melt. Provided that daily means of total discharge are well determined, the discharged volumes of groundwater and meltwater are well determined from the daily streamwater sampling. However, a slight underestimation of the discharged groundwater volumes is caused in the present study by the predominant sampling at the hours of maximum flow.

The absence of large diel fluctuations in groundwater fractions may have two explanations. Either, meltwater and groundwater are reaching the stream with constant proportions between the flows, in spite of large fluctuations in the absolute values, or, input fluctuations of tracer concentrations are damped in the channel net, including surface-water reservoirs.

In small basins, in which high flows can take place without formation of considerable surface-water reservoirs connected to the stream, attenuation of daily input fluctuations in water composition is not likely. Thus, the small fluctuations observed in Aspåsen, Buskbäcken and Svartberget show that in these basins, the groundwater fraction is rather constant during the day. The lack of fluctuations in Gårdsjön F3, on the other hand, may be due to damping of input fluctuations in tracer concentrations in the flooded bog just upstream from the sampling point. In this basin we cannot say whether there are considerable daily fluctuations in the input or not. In Nåsten and, particularly, in Stormyra the daily contributions show considerable fluctuations in calculated groundwater fractions. Considering the possibilities for damping in Stormyra, the input fluctuations in this basin, according to the present analysis, may have been still more pronounced.

8.1.5 Groundwater fractions of total discharged volumes

The discharged volumes of groundwater during the periods, given as fractions of total discharged volumes, are listed in Table 8.1 and in Table A2.1 (Appendix 2). The latter table lists the fractions calculated for different assumptions about groundwater $\delta^{18}\text{O}$, this entity either being a constant (infinite model reservoir volume) or being calculated using model reservoir volumes of 500 and 250 mm. As discussed above, the use of constant groundwater $\delta^{18}\text{O}$ yields lower limits of the groundwater fraction. The sensitivity of the calculated groundwater fractions to a systematical error in δ_p of $\pm 1\text{‰}$ and in δ_g of $\pm 0.5\text{‰}$ are given in Table A2.1 (cf. the uncertainty discussion).

For all 30 separated snowmelt events, the volume fraction of groundwater by constant groundwater $\delta^{18}\text{O}$ ranged between 32% and 95% of total flow, with a median value of 61%. For the 16 spring flood totals, the fractions ranged between 42% and 86%, median value 59%. The volume fractions of groundwater were generally largest at the comparatively small initial runoff events. Exceptions are Nåsten 1979 and Buskbäcken 1981. Using variable $\delta^{18}\text{O}$ of groundwater at the estimates, the total volume fractions increased by 3 - 6 percentage points for most of the spring floods studied.

Looking at all separated snowmelt events, the total volume fraction of groundwater decreases with increasing maximum runoff and mean rate of water input (see further Chapter 10, where the results from the snowmelt and rainfall events are all discussed together).

8.1.6 Extent of discharge areas

From the groundwater fractions of streamflow estimated by ^{18}O , basin fractions of discharge area, Y , are calculated according to Eq.(3.9):

$$Y = \frac{q}{P - \Delta S / \Delta t} (1 - X) \quad (3.9)$$

where q is specific discharge, P rate of precipitation, ΔS change in snow storage (negative during melting) over the time Δt , and X fraction of groundwater.

In the derivation of this equation all rain- or meltwater on recharge areas is assumed to infiltrate, while all such water on discharge areas is assumed to generate overland flow. The fractions of discharge area are calculated from period totals of discharged rain- or meltwater and water input, giving some sort of event mean values, see Table 8.1.

Eq.(3.9) is a very rough interpretation of the observed melt-water flow in the streams. Even if the conceptual base for the equation is disregarded, the accuracy in the calculations by Eq.(3.9) is rather low. This is true particularly for periods with small relative changes in the water equivalent of the snowpack, since ΔS then is uncertain. A prerequisite for meaningful estimates of discharge areas by Eq.(3.9) is that the total water budget is fulfilled when the whole event is regarded. For this reason, the fraction of discharge area is not given for periods with unrealistic water budgets.

In spite of the uncertainties in the assessment of Y, the estimates are considered valuable, since they provide a means for a field check of the results of hydrograph separation by ^{18}O .

For the single snowmelt events, the fractions of discharge area estimated by ^{18}O using constant groundwater $\delta^{18}\text{O}$ ranged between 2% and 60% of the basin areas, with a median value of 23%. The fraction of discharge area is positively correlated to the maximum runoff and the mean runoff of the events, see further the discussion in Chapter 10.

Field surveys of discharge areas

In order to examine whether the interpretations of the results of hydrograph separation in terms of discharge areas are reasonable, field surveys were performed during some of the spring floods. The surveys were performed either statistically or by mapping. They were carried out during one day during the main runoff events, a few days after the culmination of the spring flood.

Discharge areas were defined as areas having water on the ground or having a water table less than 5 cm below ground surface. The definition thus presupposes that saturated soil and surface-water reservoirs only occur in connection with the groundwater table. Experience from groundwater level observations in some

of the basins, and in forested till soils in general, supports this assumption. Free water in or above the ground surface very seldom occurs on areas which are not completely saturated, i.e., on areas which have an unsaturated zone above a deeper groundwater table.

The statistical surveys were carried out as spot observations made at constant intervals along straight lines. The lines were drawn to cover the whole basins. They were drawn either as equidistant, parallel lines or so as to give roughly the actual proportions of physiography and land use of the basin. Snow and water depth were also measured at the points of observation.

In the smallest basins, which have comparatively simple shapes, the discharge areas were mapped and their areas determined planimetrically.

The statistical surveys should provide upper limits of the extension of discharge areas estimated by ¹⁸O. This is so since all discharge areas are not connected to the channel network. Particularly in Nåsten which has a broken topography with a low mean-slope value, isolated discharge areas occur. In these areas surface-water reservoirs develop, which are connected to the groundwater. They drain by infiltration (and evaporation) during the successive lowering of the water table during periods of streamflow recession. Neither can all surface reservoirs which develop on outcrops be regarded as discharge areas. In elevated areas in Stormyra, small reservoirs of this type frequently develop. Some of them drain very slowly after the runoff events, by infiltration and evaporation.

The results of the field surveys are summarized in Table 8.5, showing fractions of discharge area ranging from 0.10 to 0.35 of the basins. These values are instantaneous values. If proportionality is assumed between the extent of discharge areas in a basin and streamflow, the event maximum and mean values are obtained by multiplying the measured fractions by $q_{\max}/q_{\text{day}}$ and $q_{\text{mean}}/q_{\text{day}}$ respectively (see these ratios given in Table 8.5). For Stormyra the maximum fraction obtained is >1 and thus unreasonable. This might be an effect of the geology and topography of this basin. Flat areas with loose deposits are here often surrounded by steep, bare rocks. When the discharge areas in the loose deposits reach the bordering bedrock, they cannot expand any more. For the other basins the estimated maximum fractions are not unrealistic, but the figure seems high for Buskbäcken (0.70) and, considering the topography of the basin,

for Gårdsjön F3 (0.45). According to similar comparisons with the mean values, the event mean fractions generally should be slightly lower than the observed ones.

Table 8.5 Field surveys of discharge areas.

a. Experimental situations and observed fractions of discharge areas.

Basin	Date	Number of observations	Fraction of discharge area (%)
Aspåsen	17/5 -81	mapping	≈10
Buskbäcken	26/4 -80	602	28
Gårdsjön F3	12/4 -80	mapping	15
Nåsten	6/4 -79	544	23
	3/4 -82	1181	31
Stormyra	6/4 -82	553	35
Svartberget Nedre	14/5 -81	201	25

b. Hydrologic situation.

	q_{day} ($l s^{-1} km^{-2}$)	q_{max}/q_{day}	q_{mean}/q_{day}	surface water storage (mm)	snow-covered area fraction of the basin (%)	water equivalent of snow storage (mm)
Aspåsen	107	2.2	1.1		<40	50
Buskbäcken	42	2.5	0.9	4.0	22	11
Gårdsjön F3	25	3.0	0.9		<10	<5
Nåsten	36	1.3	0.7	14.0	38	19
	72	1.3	0.6	12.9	49	29
Stormyra	37	5.3	0.8	6.0	26	14
Svartberget Nedre	81	1.4	0.8		30	23

q_{day} = mean specific discharge of the day

q_{max} = maximum specific discharge of the event

q_{mean} = mean specific discharge of the event

The largest volume of surface water was found in Nåsten. This finding is reasonable considering the physiographical characteristics of this basin, as compared to the others. Somewhat

surprisingly, the surface water storage in this basin was about the same in the two surveys although the stream discharge was twice as large during the second survey.

The fraction of discharge area obtained by ^{18}O and that obtained by field survey are compared in Fig.8.8. Basin fractions by ^{18}O are given both for constant $\delta^{18}\text{O}$ of groundwater (infinite reservoir) and for the reservoir volume of 250 or 500 mm, providing the best fit of groundwater $\delta^{18}\text{O}$ to that of streamwater after the event.

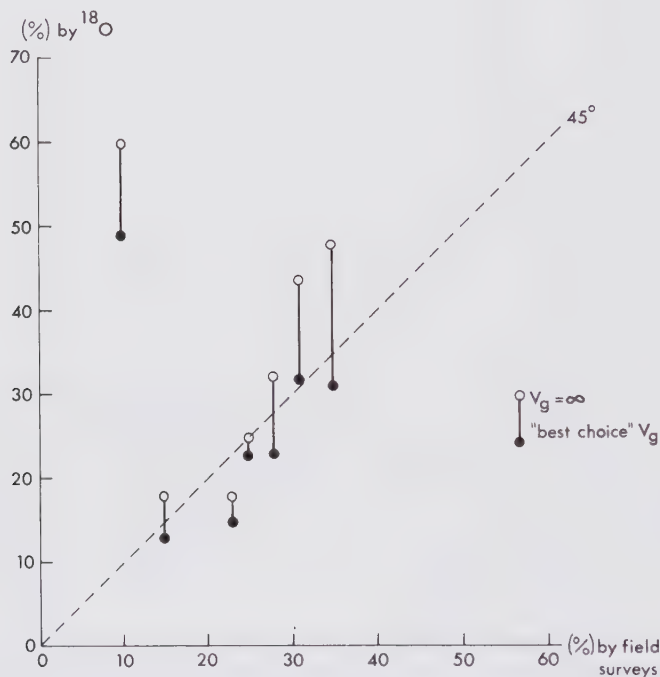


Fig.8.8 Basin fraction of discharge area estimated by ^{18}O and by field surveys. The "best-choice" value is calculated by using the best fitting model reservoir volume of 250 or 500 mm (see text).

The observations from Aspåsen differ greatly by any method, having a five-times larger discharge area by ^{18}O than was observed in field. For the other basins the agreement must be considered good.

8.1.7 Conclusions and comments on snowmelt events basin by basin

Aspåsen

The investigated spring flood was caused by melting of the largest snowpack encountered in this study. The main runoff event showed the smallest volume fraction of groundwater of the investigated melting events, 32% by constant $\delta^{18}\text{O}$ of groundwater. The calculated fraction of discharge area for this event, 60%, differed considerably from the one obtained by a field survey, 10%.

Buskbäcken

Hydrograph separation was performed for two spring floods of similar, normal, magnitudes. Large snowfalls, drastically changing the $\delta^{18}\text{O}$ of snowpack, appeared at the end of the main runoff events, but hydrograph separation remained possible. A third spring flood, in 1982, was also sampled. But unfavourable precipitation and unreliable discharge values prevented hydrograph separation.

The total volume fractions of groundwater and calculated basin fractions of discharge area were about the same both years. One field survey of discharge areas agreed acceptably with the calculated value during the related event.

Small diel fluctuations in groundwater fractions were found in 1980, but in 1981 no fluctuations were seen. Reasons for the difference between the two years could not be given.

Gårdsjön F1, F2, and F3

The investigated spring flood showed a clear dominance of groundwater in the three basins. A rainfall of 13 mm occurred at the onset of melting, when the water equivalents of the snowpacks were around 40 mm. Fortunately the $\delta^{18}\text{O}$ of the rainwater was close to that of snowpack, so it introduced little error in the hydrograph separations.

Due to an unrealistic water budget (cf. the residual term r in Table A2.1), the fraction of discharge area could not be calculated for basin F1. In basin F3, a field survey gave a fraction of discharge area close to the one obtained by ^{18}O .

Diel fluctuations in groundwater fractions were not seen in the intensively sampled basin (F3). Attenuation of $\delta^{18}\text{O}$ -fluctuations of the water input to the stream cannot, however, be excluded in this basin.

Nåsten

Two spring floods, one moderate and one large, were separated. Sampling was performed also during a third, small spring flood, but the isotopic conditions did not allow separation, except for a very small initial event. Due to uncertain discharge data, summation was not performed for this event.

During the first days of melting, a negative peak in streamwater $\delta^{18}\text{O}$, indicating a peak in the relative contribution of meltwater, was often found. The negative peak of streamwater $\delta^{18}\text{O}$ in April 1980 probably resulted from rainfall and snowfall with low δ -values. For certain parts of the large spring flood in 1982, separation is uncertain due to the influence of precipitation being heavier in ^{18}O than the snowpack. During such periods the groundwater fraction tends to be overestimated.

Field surveys of discharge areas agreed roughly with estimates from ^{18}O .

Stormyra

The same three spring floods as those in Nåsten were investigated and as in Nåsten only two could be separated. Here too, the details of the separation in 1982 are uncertain due to precipitation. The very high groundwater fraction in Stormyra I 1979 is accurately determined, since the isotopic difference between streamwater before melting and snowpack was large and the amount of precipitation was small. The groundwater flow during the first part of Stormyra II 1979 might be overestimated, however. The considerable amount of sleet (unfortunately not sampled for ^{18}O) occurring then may, if it was higher in ^{18}O than the snowpack, have been partly responsible for the high δ -values of streamwater and thus for the large calculated groundwater flow during the hydrograph rise.

The fraction of discharge area established by the field survey in 1982, 0.35, was the largest among those observed in this study.

Svartberget

Two spring floods of similar, normal sizes were separated at three sampling points: Svartberget Övre, Svartberget Västra, and Svartberget Nedre. Accurate discharge values were available only at Svartberget Nedre. When calculating the period means of groundwater fraction, the discharge values from this station were used for Svartberget Övre in 1981 and for Svartberget Västra for both years. In this way, the instantaneous groundwater fractions at the ungauged stations were weighted against discharge, assuming the same shape of the hydrographs at the stations. Since the absolute values of discharge at the ungauged stations are not known, the fraction of discharge area could be calculated only for Svartberget Nedre.

In 1981 there was a considerable difference in groundwater fractions at Svartberget Övre and Svartberget Västra, the former having much lower fractions. The large contribution of meltwater at Svartberget Övre is probably due to the mire, representing a large area for saturated overland flow production. The influence of the mire is also seen at Svartberget Nedre, where the water is to a large degree a mixture between water from the two sub-basins. In 1982 the difference between Västra and Övre remains, but now less prominent.

Rainfall of high ^{18}O content makes the hydrograph separations uncertain during the second half of period II 1982.

A field survey of discharge areas in 1981 agreed satisfactorily with the results obtained by ^{18}O .

8.2 Rainfall events

A prerequisite for hydrograph separation by ^{18}O is that the $\delta^{18}\text{O}$ of water input (meltwater or rainwater) differs markedly from that of the groundwater. Thanks to the seasonal variation in precipitation $\delta^{18}\text{O}$, this condition was mostly fulfilled during the investigated spring floods. The snowpacks, roughly representing the weighted mean of winter precipitation $\delta^{18}\text{O}$, were depleted in ^{18}O as compared to the streamwater before melting.

At rainfall-generated runoff events the isotopic conditions may or may not allow separation. Even if summer precipitation in its mean values is enriched in ^{18}O as compared to the groundwater, there is a large variability in $\delta^{18}\text{O}$ within and among single precipitation events (cf. Figs. A1.1-9).

A rainfall-generated runoff event which is ideal for separation by ^{18}O should have the following characteristics:

1. The rainfall is preceded by a period of dry weather to ensure pure groundwater flow before the event and further to minimize the contribution of stored surface water to the stormflow.
2. The $\delta^{18}\text{O}$ of rainfall differs markedly from that of streamwater before the event.
3. The variability in rainfall $\delta^{18}\text{O}$ over time is small both within and among rainfalls associated with the event.

In this study, hydrograph separation by ^{18}O could be performed for fifteen rainfall-generated runoff events. For a further four events, valuable information could be obtained from the isotopic data, although complete hydrograph separation was not possible. The rainfall events which could be separated by ^{18}O , or which could give valuable information, are marked in Figs. A1.1-9. As for the spring floods, the period boundaries for each event are not strictly defined from the hydrographs. The boundaries are also chosen to enclose periods for which isotopic separation could be performed, thus taking the isotopic situation and the data available into consideration.

The duration of rainfall-generated runoff events in the investigated basins is typically 3-15 days. Of the sampled events, the longest duration was in Nâsten after a large autumn rain.

The shortest duration was observed in Aspåsen after a small early autumn rain.

As mentioned in Chapter 5, streamwater was sampled at various intensities. A few runoff events were sampled several times a day. Many events were sampled daily and some at larger time intervals.

Rainwater was sampled once a day, except for the event in Aspåsen and those in the Gårdsjön basins, where denser rainwater sampling was performed.

As with the procedure during the spring floods, the $\delta^{18}\text{O}$ of groundwater, δ_g , was assumed equal to that of streamwater before rainfall. Modifications of this assumption, necessitated by changes in streamwater $\delta^{18}\text{O}$ before the separated event, are described separately for each event. For a few events, which had complete data and favourable isotopic situations, separation could be performed also with variable groundwater $\delta^{18}\text{O}$ (Eq.3.6). But for most events the calculations were made using constant groundwater $\delta^{18}\text{O}$ only. As $\delta^{18}\text{O}$ of rainwater, δ_p , a weighted mean of the samples associated with the events was used, or accumulated weighted means, according to procedures described separately for each event. The occurrence of surface water before the events can be judged from the hydrographs in Figs.A1.1-9. The smaller the discharge, the smaller the surface reservoirs.

Quantitative results of the separations, i.e., the total volume fraction of groundwater and the basin fraction of discharge area, are given in Table 8.6 together with some hydrologic characteristics of the events. Further results, and more detailed characteristics are given in Appendix 3, Table A3.1. Due to the large variability in the hydrologic and isotopic conditions and in the amount of data available, the accuracy of the calculated fractions varies greatly.

As pointed out earlier, the investigated rainfall events differ from the snowmelt events in that the isotopic situations, as well as the amount of data available, vary considerably among different events. In order to make possible an evaluation of the applied method and of the quantitative results, it is therefore considered most suitable to present the rainfall events individually.

Table 8.6 Results of hydrograph separation: rainfall events

Basin	Event	Maximum discharge ($1 \text{ s}^{-1} \text{ km}^{-2}$)	Total runoff (mm)	Fraction of groundwater (%)	Fraction of discharge area (%)
Aspåsen	A	80	7.9	87	7
Buskbäcken	A	219	31		
Gårdsjön	F1 A	111	13.6	81	4
	F1 B	11	1.5	96	0.4
Gårdsjön	F3 A	90	8.5	81	2.5
	F3 B	16	1.7	87	1.3
Nåsten	A	98	41	81	17
	B	32	13	78	8
	C	11	3.6	≈99	0
	D	15	5.6	>66	<4
	E	2.2	0.60	95	0.2
Stormyra	A	19	5.3	81	2.1
	B	6	2.9	86	1.8
	C	145	21.5	≈68	≈15
	D I	11	2.3	>63	<3.7
	D II	120	24.5	>53	<25
Svartberget	A	54	11.1	92	3.0
Nedre	B	22	4.1	85	3.5
	C	30	16.2	99	0.4

For each event a diagram is presented showing total hydrograph and calculated groundwater hydrograph, streamwater $\delta^{18}\text{O}$, rainwater $\delta^{18}\text{O}$, and amount of rainfall. For some events, the rainwater hydrograph is also shown. In drawing the total hydrographs for events with sparse streamwater sampling, data from water level recorders have been used. For other events, the total hydrographs are based on the observed water level at the moments of streamwater sampling.

Events for which the most complete data are available, i.e., Gårdsjön A and B, and also Stormyra A and B, are described first in some detail. Thereafter two events with "inexplicable" variations in streamwater $\delta^{18}\text{O}$ are discussed, Buskbäcken A and Svartberget Nedre C. The remaining events are presented with a few comments in Appendix 3, with the comments focusing mainly

on the assumptions made and on the problems arising in interpretation.

For some of the intensively sampled events conclusions could be drawn on the time variation of the groundwater fraction. These results were helpful in the attempts to separate the less intensively sampled events, for which total volume fraction of groundwater could be estimated or lower limits be determined.

After the presentation of the events, some common features of the variation in the groundwater and rainwater fractions within the events are pointed out. The obtained total volume fractions of groundwater and basin fractions of discharge area are discussed and related to flow conditions in Chapter 10, together with those of the snowmelt events.

8.2.1 Intensively sampled events

Gårdsjön F1 A, June 1982 (Figs.8.9a and 8.9b)

After a long dry period a frontal rain started at noon on June 12. During 36 h of continuous rainfall a total amount of 56 mm was recorded and sampled for ^{18}O at time intervals of 1-3 h. Maximum 15 min intensity was 6 mm h^{-1} .

As is seen in the figures, the $\delta^{18}\text{O}$ of rainwater varied considerably during the rainfall. The first 3.9 mm was isotopically heavier than the streamwater, whereas the bulk volume of rainwater was isotopically considerably lighter than the streamwater.

Due to the time lag between rainfall input and streamflow output, it was not considered possible to utilize the varying $\delta^{18}\text{O}$ of rainwater when carrying out the separation by Eq.(3.3). The separation was performed using the weighted mean $\delta^{18}\text{O}$ of rainfall, -13.6‰ , as δ_p . By using the mean value in the calculations the variation in the groundwater fraction, and thus the groundwater and rainwater hydrographs, is not accurately determined.

In this particular case the total volume fraction of groundwater is probably underestimated by the use of the mean rainfall $\delta^{18}\text{O}$. This is so since for a constant (or increasing) rainfall intensity, the disposition of the ground to generate overland flow (Hortonian as well as saturation) generally increases as the rainfall proceeds, due to increasing soil

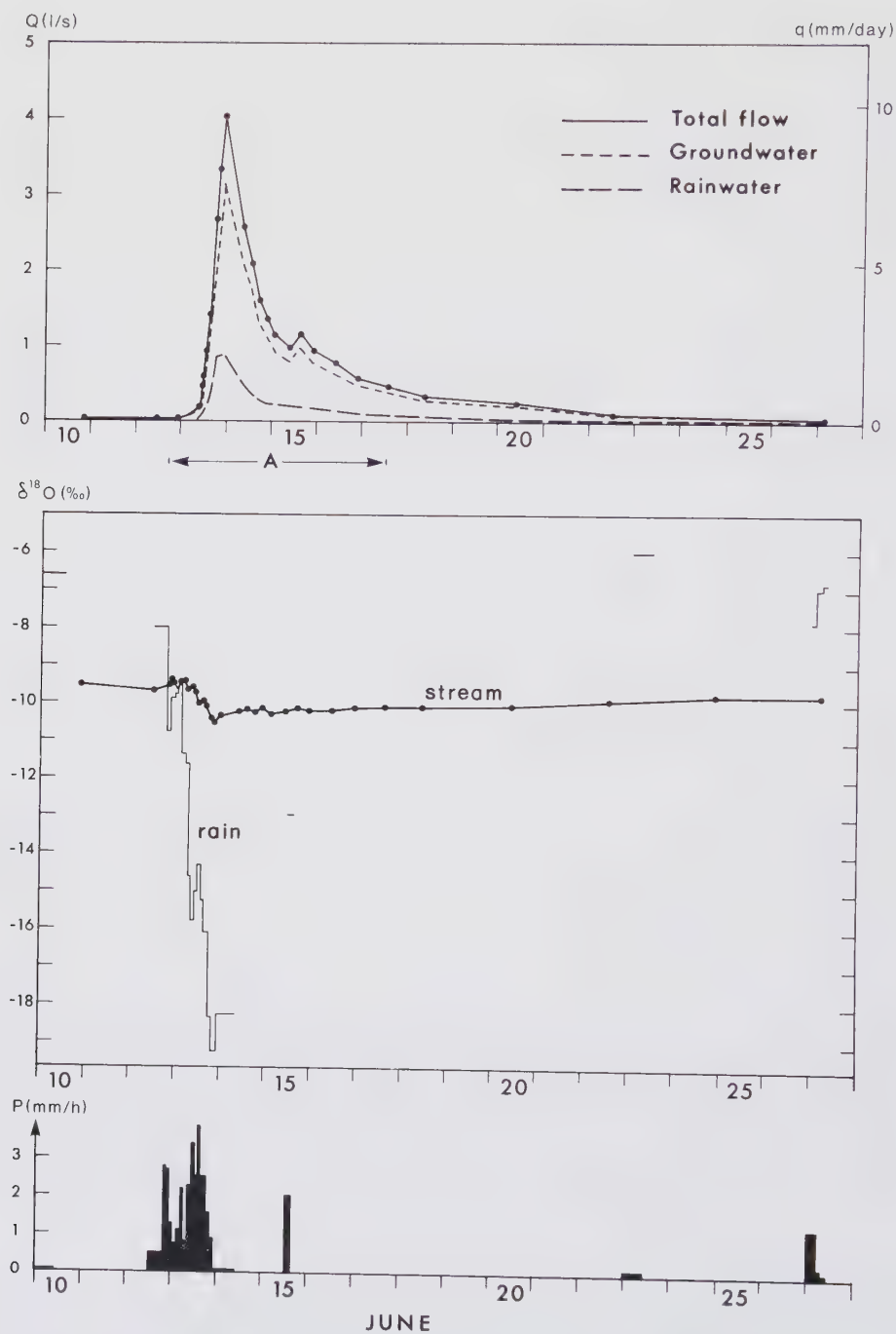


Fig.8.9a Stream hydrograph with calculated flows of groundwater and rainwater, $\delta^{18}O$ of streamwater and rainwater, and rain intensity. Gårdsjön F1 A, June 1982.

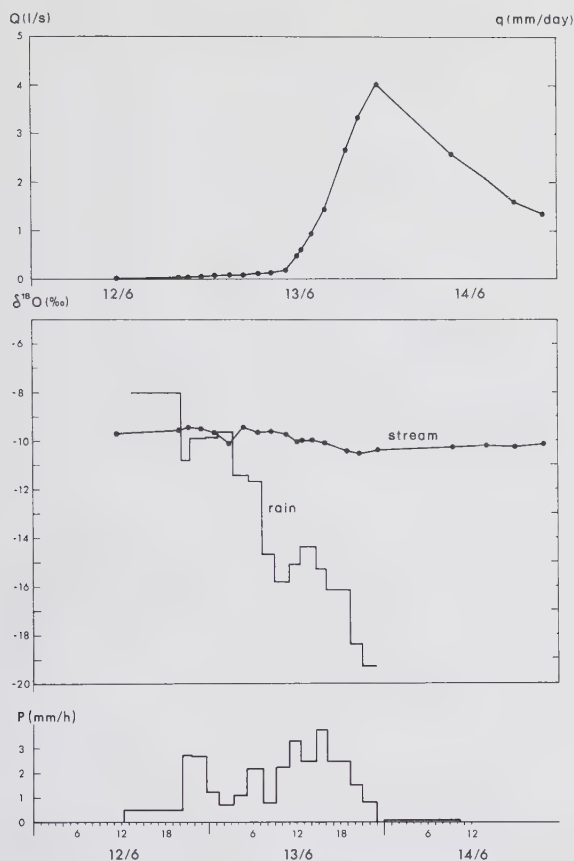


Fig.8.9b Stream hydrograph, $\delta^{18}\text{O}$ of rainwater and streamwater, and rain intensity. Enlarged time scale. Gårdjön Fl A, June 1982

moisture and expanding saturated areas. The rainwater contribution to streamflow is therefore likely to increase during the rainfall. Since the $\delta^{18}\text{O}$ of rainwater in this storm, with the exception of the first few millimetres, showed a growing deviation from that of groundwater, the δ -value of the rainwater actually contributing to the streamflow probably deviated more than the mean value, which was used as δ_p in the calculations.

A convective shower of 5.3 mm occurred on June 15, having a maximum 15 min intensity of about 15 mm h^{-1} . Fortunately, its $\delta^{18}\text{O}$ did not differ much from the mean value of the main storm. Consequently it had little effect on the hydrograph separation, which was done using the δ -value of the main storm as δ_p during the whole runoff event. The shower resulted in a small secondary runoff peak, but no effect on streamwater $\delta^{18}\text{O}$ was observed, indicating that the flow increment was mainly by groundwater in

this case as well. It should be noted, however, that the sampling intensity in connection with this shower was too low to allow a detailed analysis of the flow components.

From June 12 to June 17,¹ a total volume corresponding to 13.6 mm was discharged. According to the separations by ^{18}O , 18% of this volume constituted new rainwater while the remainder, 82%, was groundwater. The smallest instantaneous value of the groundwater fraction, about 0.75, seems to have occurred just before or at the time of peakflow, but the details of the separation are uncertain due to the varying rainfall $\delta^{18}\text{O}$, as discussed above. The volume of rainwater discharged up to June 17 corresponds to the precipitation over 4% of the basin area, which area fraction is interpreted as the basin fraction of discharge area (cf. Eq.3.9).

The fact that streamwater $\delta^{18}\text{O}$ did not return to its original value after the runoff event is probably due to changed groundwater $\delta^{18}\text{O}$ by infiltration during the rain. Using variable groundwater $\delta^{18}\text{O}$ by Eq.(3.6), a model reservoir volume of 200 mm describes the observed total change in streamwater $\delta^{18}\text{O}$. The corresponding total fraction of groundwater was 85%.

Gårdsjön F1 B, July 1982 (Fig.8.10)

A second runoff event suitable for separation by ^{18}O occurred in early July. After 17 days with little rainfall, 16.6 mm of rain fell on July 3, having a maximum 15 min intensity of 16 mm h⁻¹. (Note the enlarged discharge scale in Fig.8.10 as compared to Fig.8.9.) This time, the $\delta^{18}\text{O}$ of the rainwater was higher than that of the streamwater, and the in-storm variations were small. Streamwater $\delta^{18}\text{O}$ deviated from its original value only during the rainfall and the first few hours after it.

The calculated groundwater fraction of the total flow volume between July 3 and July 6 was 96%, with a minimum instantaneous fraction of about 0.85 just before peakflow. The fraction of discharge area according to Eq.(3.9) was 0.03.

¹ In order to facilitate comparison with the less intensively sampled stream F3, June 17 was chosen as the end of the period. From this day to the end of runoff event around June 27 about 3 mm were discharged, i.e., 20% of the total event runoff. The calculated groundwater fraction continuously increased during this late part of the event.

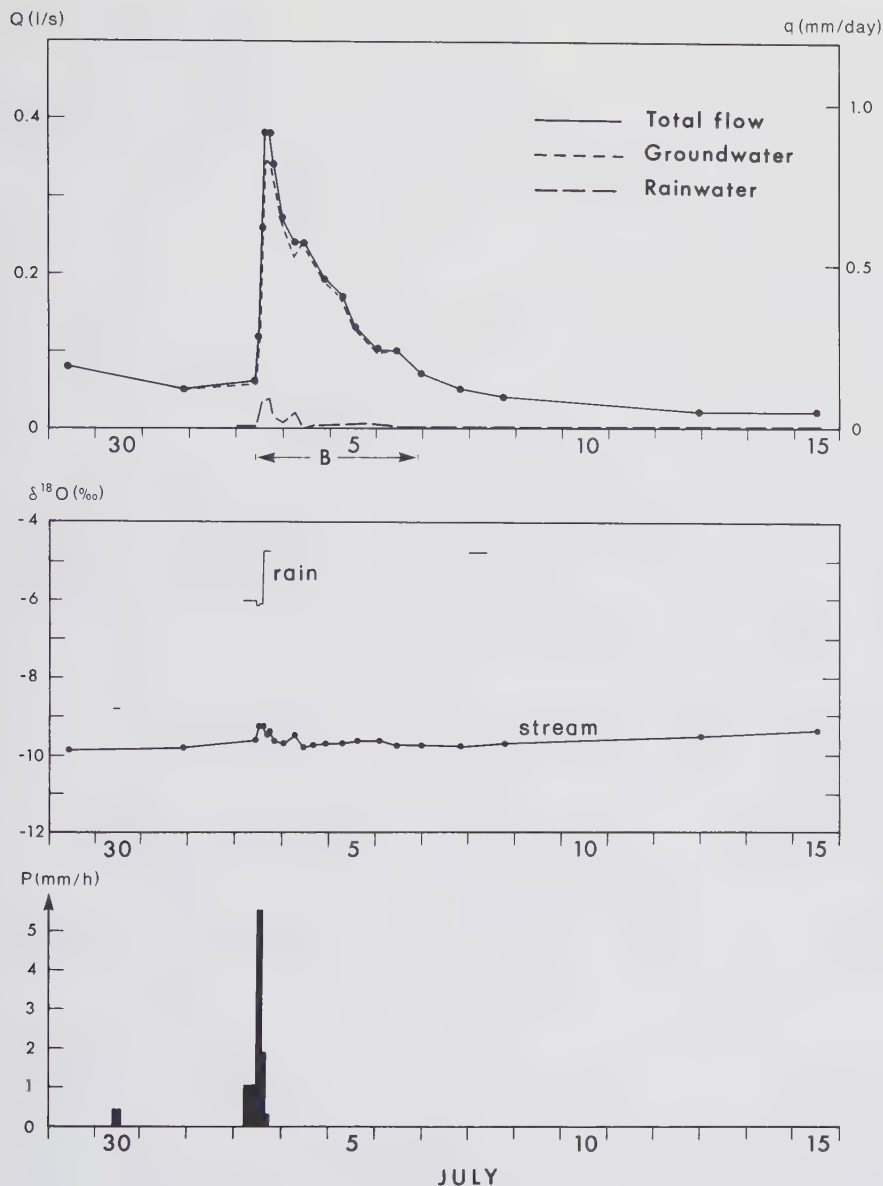


Fig.8.10 Stream hydrograph with calculated flows of groundwater and rainwater, $\delta^{18}O$ of streamwater and rainwater, and rain intensity. Gårdsjön B, June - July 1982.

Thanks to the comparatively constant rainfall $\delta^{18}O$, the calculated variation of the flow components is more accurate in this case than in the June event. This is so also since the groundwater $\delta^{18}O$ does not seem to have changed during the event, which is reflected by the return of streamwater $\delta^{18}O$ to its

pre-storm value after the event. The different behaviour of the $\delta^{18}\text{O}$ of the contributing groundwater at the June and July events is quite natural, considering the much smaller rainfall and runoff volumes in the latter event. A corresponding difference between the events was also observed in the $\delta^{18}\text{O}$ of groundwater samples from observation wells in the basin.

Gårdsjön F3 A and B

The above rainstorms generated hydrographs in stream F3 with shapes similar to those in stream F1. The isotopic content also varied in a similar way (Fig.8.11), showing little contribution of rainwater.

The calculated groundwater fractions in the June event are very similar in the two basins, but in the July event the isotopic data gave a slightly lower groundwater fraction in stream F3 (Table 8.6). Because of the large difference between δ_g and δ_p in the two storms, the calculated groundwater fractions are not very sensitive to errors in δ_p .

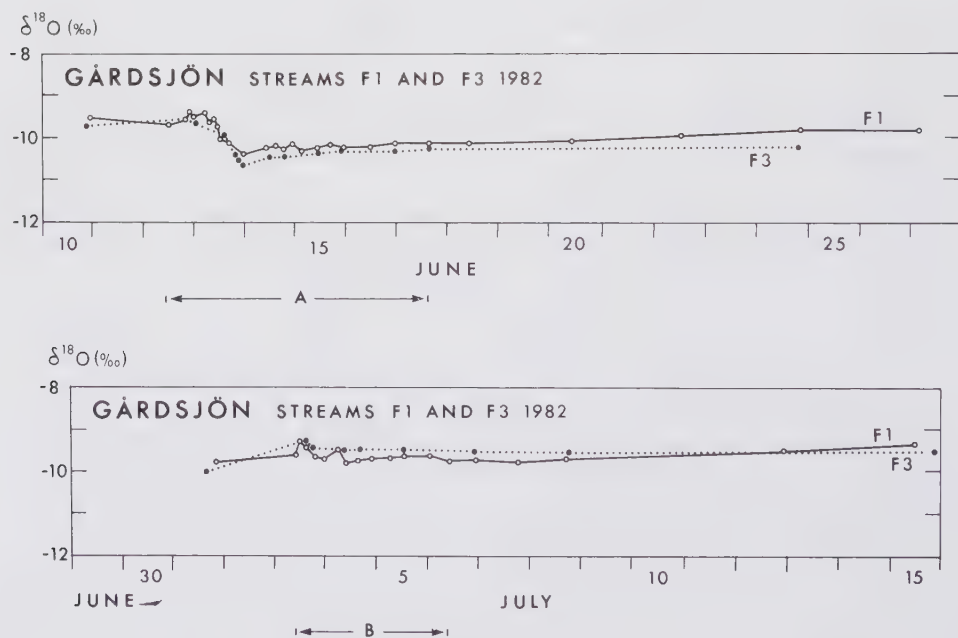


Fig.8.11 $\delta^{18}\text{O}$ of streamwater. Gårdsjön F1 and F3, events A and B, June and July 1982.

Stormyra A, August 1979 (Fig.8.12)

As for the Gårdsjön events, the hydrologic conditions for hydrograph separation were good at this intensively sampled event, with an initially comparatively dry basin and mainly one rainfall. Unfortunately the variation in rainfall $\delta^{18}\text{O}$ prevented a very accurate separation by ^{18}O . But since the first two rainfall samples, accounting for 42 mm out of the 47 mm of rainfall associated with the main discharge peak, were isotopically not too different and on the same side of the groundwater $\delta^{18}\text{O}$, a separation could be performed.

As daily values of δ_p the accumulated weighted mean of rainfall $\delta^{18}\text{O}$ were used, calculated separately for the two groups of rainy days. As δ_g the $\delta^{18}\text{O}$ of streamwater on August 20 was chosen, after which day a rise in streamwater $\delta^{18}\text{O}$ occurred.

According to the isotopic separation, rainwater dominated the hydrograph during the early rise, whereas the peakflow and recession were dominated by groundwater. Direct rainwater contribution to the second peak seems to have been very small. Of the total discharged volume of the main peak, i.e., up to August 31, the groundwater fraction was estimated at about 80%.

Even if the exact shapes of the rainwater and groundwater hydrographs are uncertain, their internal relationship is reasonable, with the groundwater hydrograph being delayed in time and having the smoothest peak. It should be noted that the smallest instantaneous groundwater fraction, 0.07, appeared when the discharge still was low, only a few l s^{-1} .

Stormyra B, October 1979 (Fig.8.13)

The event was preceded by dry weather. With the exception of 0.8 mm of rainfall sampled over October 17, the daily values of rainfall $\delta^{18}\text{O}$ associated with the first peak showed small variations as compared to the difference between rainwater and groundwater $\delta^{18}\text{O}$. Hydrologic and isotopic conditions for hydrograph separation thus were good. Unfortunately, the rainfall generating the second peak was not correctly sampled, so separation could be done only up to October 19.

Streamflow was dominated by groundwater on all sampling occasions. The largest relative contribution of rainwater, 0.30, was found during the later part of the hydrograph rise.

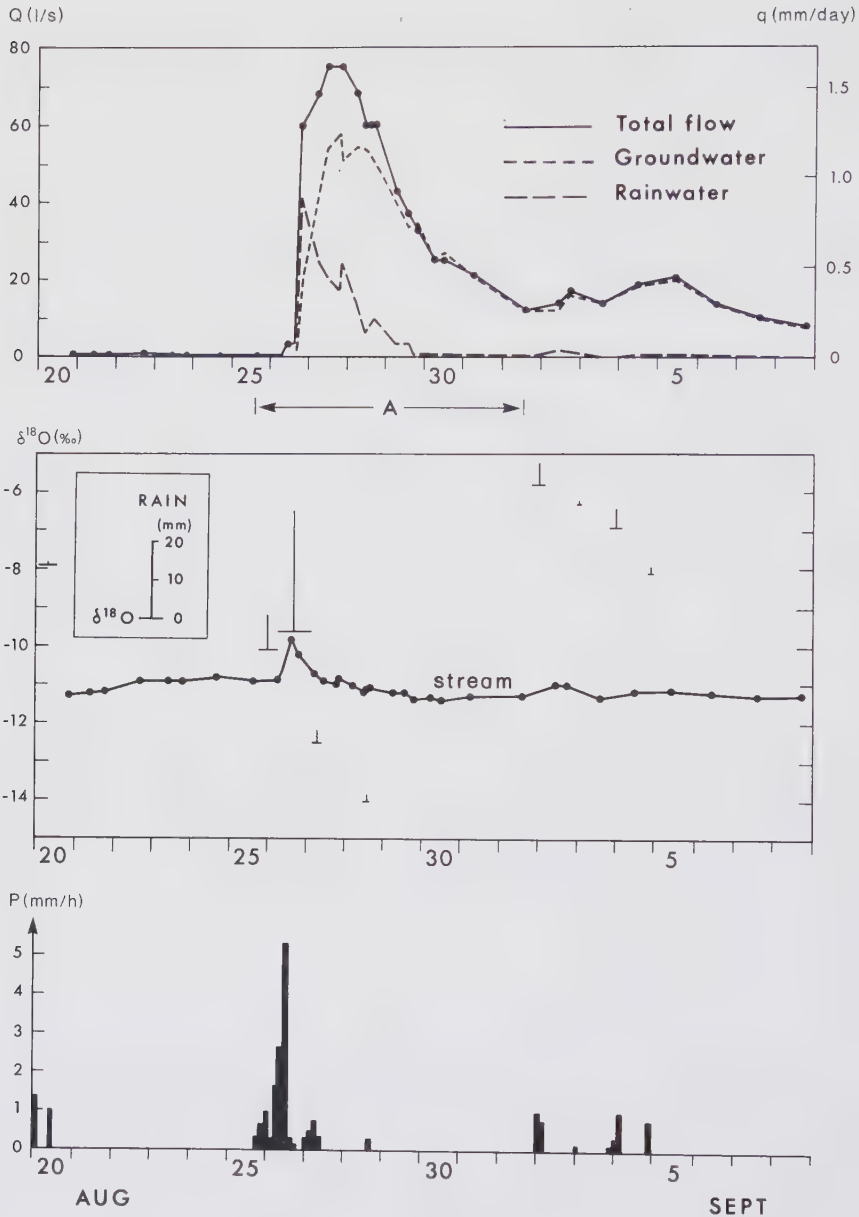


Fig.8.12 Stream hydrograph with calculated flows of groundwater and rainwater, $\delta^{18}O$ of streamwater and rainwater, and rain intensity. The level of the horizontal lines in the $\delta^{18}O$ diagram marks the $\delta^{18}O$ value of the rainfall samples and the length of these lines the duration of the rainfall. The vertical lines show the amount of rainfall (see scale). Stormyra A, August - September 1979.

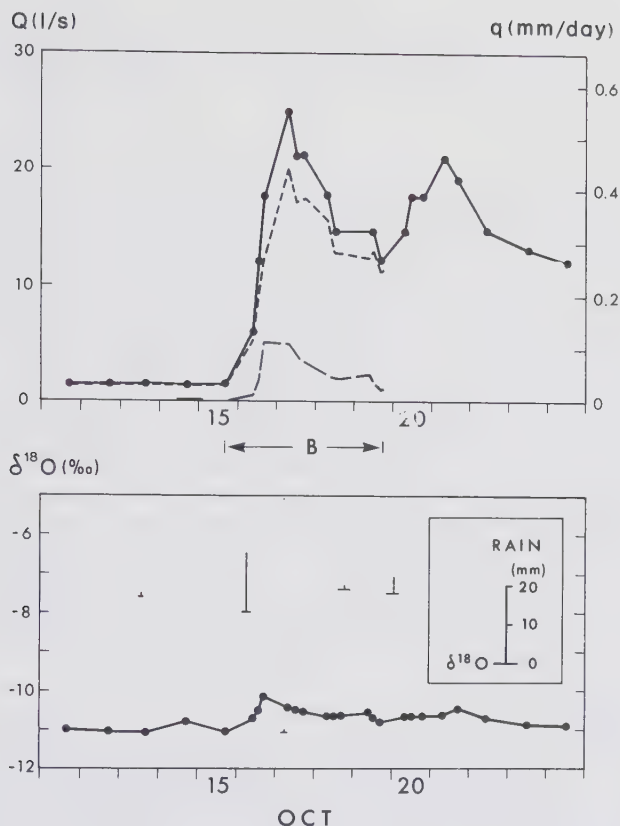


Fig.8.13 Stream hydrograph with calculated flows of groundwater and rainwater, $\delta^{18}O$ of streamwater, and amount and $\delta^{18}O$ of rainfalls. Stormyra B, October 1979. Legend see Fig.8.12.

There are, however, too few samples around the time of maximum flow to give very accurate shapes to the rainwater and groundwater hydrographs. With the data available the latter seems to be a little delayed. Of the total volume discharged during October 15-19, the groundwater fraction was estimated at 86%.

Methodological conclusions from the intensively sampled events

Rainfall $\delta^{18}O$

The results from Gårdsjön show that there may be large variations in rainwater $\delta^{18}O$ within a rainfall (Gårdsjön A). Other rainfalls may have a more constant isotopic composition over time (Gårdsjön B). In the case of Gårdsjön A, the changes in rainfall $\delta^{18}O$ were such that the use of the weighted mean of

rainfall $\delta^{18}\text{O}$ gave a satisfactory accuracy as to total discharged volumes of groundwater and rainwater. The shapes of the respective hydrographs were, however, less accurately determined.

During other rainfalls, the in-storm variations of rainwater $\delta^{18}\text{O}$ may be such that hydrograph separation by the present model would be impossible. Since the ability of rainfall to generate overland flow increases as the rainfall proceeds, and also since there is a time lag between rainwater input and streamflow output, the use of the weighted mean might give erroneous results.

Except for Aspåsen A and Gårdsjön A and B, all separations in this study are based upon daily rainfall samples. It is thus evident that the results of some separations might be incorrect due to unfavourable in-storm variations in rainfall $\delta^{18}\text{O}$. But since the influence of these variations on the separations probably is random, there is no reason to distrust the general tendencies found in the several separations presented here.

Streamwater $\delta^{18}\text{O}$

The largest rainwater fractions seem to be found during the hydrograph rise. After peakflow, streamwater $\delta^{18}\text{O}$ is either constant or tends to return to its pre-storm value, showing constant or increasing groundwater fractions. In no case, either in the examples presented above or in those presented below and in Appendix 3, did the data indicate a decrease in groundwater fraction after peakflow.

The graphs of streamwater $\delta^{18}\text{O}$ versus time are comparatively smooth, indicating that valuable information can also be obtained from sparser streamwater sampling than in the above examples. A good determination of the shapes of the rainwater and groundwater hydrographs needs several samples during the hydrograph rise. But rough estimates of total volume fractions of groundwater and rainwater can be based upon comparatively sparse sampling, provided that there is one sample taken close to the time of peakflow.

8.2.2 Events with "inexplicable" variations of streamwater $\delta^{18}\text{O}$

Buskbäcken A, June 1980 (Fig.8.14)

This intensively sampled event could not be separated due to unfavourable isotopic conditions, but the results are still considered to be of interest. After four days with isotopically heavy rainfall ($\Sigma P = 21 \text{ mm}$, $\delta_p = -9,5\text{‰}$), wetting the basin and causing a small discharge peak, a large rainstorm entered the basin on June 23. A rainfall of 45 mm on June 23, isotopically depleted as compared to the initial streamwater, generated a large discharge peak (maximum flow being almost twice that of the spring floods of 1980 and 1981). The hydrograph recession was interrupted on the afternoon of June 24 by 6 mm of rainfall being isotopically heavier than the streamwater. The rainfalls associated with the second peak, around June 27, were, on the other hand, all depleted as compared to the initial streamflow.

There are no streamwater samples available between June 18 and 23, i.e., during the small "pre-event" discharge peak. During the main peak, streamwater $\delta^{18}\text{O}$ seems to have increased, at least as compared to the initial streamwater sample from June 18. Clearly, streamwater during this peak cannot be a mixture of rainwater with $\delta^{18}\text{O}$ equal to the weighted mean and of groundwater with $\delta^{18}\text{O}$ equal to the initial streamwater. Reasons for the apparent increase in the $\delta^{18}\text{O}$ of streamwater during the main peak might be:

1. Influence of surface water from the pre-event isotopically heavy rainfall. This is not probable, in view of the small discharge, and thus small surface-water reservoirs, during the "pre-event" peak as compared to the main peak.
2. Influence of soil water and groundwater recharged during the pre-event rainfalls. The high δ -values could be an effect of increasing groundwater $\delta^{18}\text{O}$. Isotopically heavy soil water is then assumed to be transformed into groundwater and discharged into the stream by the main, isotopically light, rainfall. Since any change in groundwater $\delta^{18}\text{O}$ is unlikely to exceed, say, $0.5\text{--}1.0\text{‰}$, this interpretation would give a dominance of groundwater during peakflow. The decrease in streamwater $\delta^{18}\text{O}$ after June 25 could in a similar way be explained by a contribution of groundwater having infiltrated during the main rainfall.

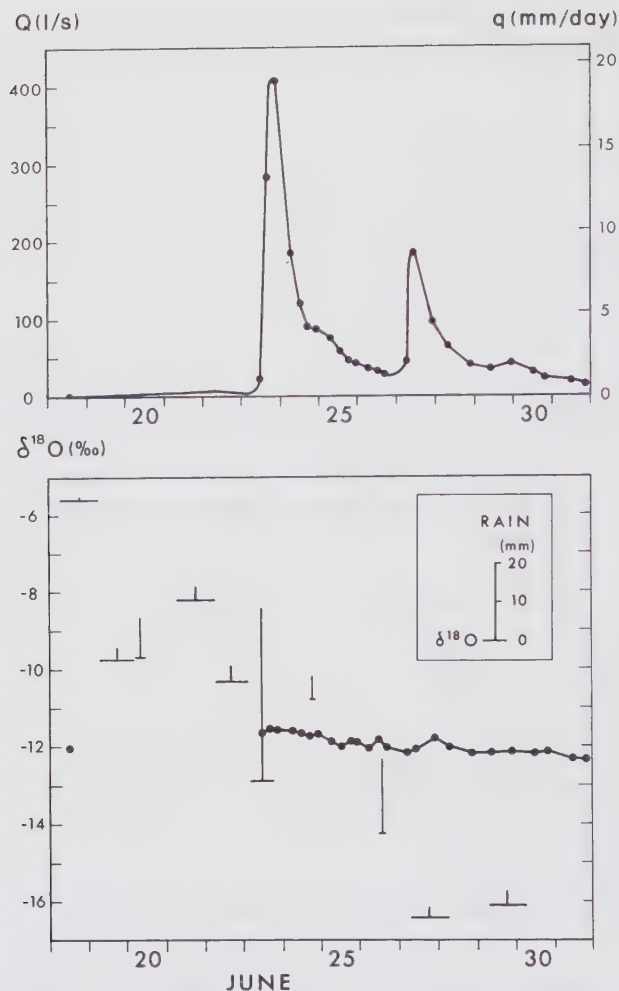


Fig.8.14 Stream hydrograph, $\delta^{18}\text{O}$ of streamwater, and amount and $\delta^{18}\text{O}$ of rainfalls. The length of the horizontal rainfall lines shows the duration of rainfall when known; if the duration is not known the length marks the 24 h sampling period. Buskbäcken A, June 1980.

3. Large increase in rainwater $\delta^{18}\text{O}$ with time during the main rainfall (cf. Gårdsjön A). It is thereby assumed that the first part of the rainfall mainly wets the soil, while a certain amount of the last part, being higher in ^{18}O than the initial streamwater, contributes directly to streamflow. Since the actual in-storm variation of rainfall $\delta^{18}\text{O}$ is not known, the relative contributions of rainwater and groundwater cannot be quantified.

The second runoff event, around June 27, seems to have been completely dominated by groundwater, since streamwater $\delta^{18}\text{O}$ does not change significantly although the rainwater is considerably lower in ^{18}O than the initial streamwater. The wet basin, as reflected by the large discharge before the peak, and the large isotopic difference between rainfall and initial streamwater, excludes false conclusions resulting from changing rainfall $\delta^{18}\text{O}$.

It is concluded that quantitative hydrograph separation of this event is not possible. Unless the hydrological and isotopic conditions were somewhat extreme, the runoff event was dominated by groundwater.

Svartberget Nedre A, August 1981 (Fig.8.15)

The isotopic changes of streamwater during this event are somewhat surprising. Although the bulk of the rainfall was considerably lower in ^{18}O than the initial streamwater, streamwater $\delta^{18}\text{O}$ increased. During recession it decreased, eventually reaching the initial value. The δ -values of the rainwater samples have been rechecked. The anomalous isotopic behaviour of the streamwater is similar to that found at Buskbäcken A, but the isotopic difference between initial streamwater and rainwater is larger, and the variation in streamwater $\delta^{18}\text{O}$ is better determined. For Buskbäcken, the following possible reasons for the unexpected behaviour of streamwater $\delta^{18}\text{O}$ were discussed:

1. Influence of pre-event surface water.
2. Influence of soil water and groundwater recharged during pre-event rainfalls.
3. Large increase in rainwater $\delta^{18}\text{O}$ during rainfall.

As for Buskbäcken, the first point is not probable for Svartberget Nedre A. The second point cannot be excluded. Isotopic data from the rainfalls prior to those shown in Fig.8.15 are not available, but high values of streamwater $\delta^{18}\text{O}$ observed in late July support this explanation. Considering the expected magnitudes of changes in groundwater $\delta^{18}\text{O}$, the explanation implies that the runoff event was heavily dominated by groundwater. The third point is unlikely to explain the isotopic changes found at Svartberget Nedre A, since the difference

between initial streamwater and mean rainfall $\delta^{18}\text{O}$ is very large. Further, the observed difference between the two rainwater samples would rather suggest a decrease in rainwater $\delta^{18}\text{O}$ during the rainfall.

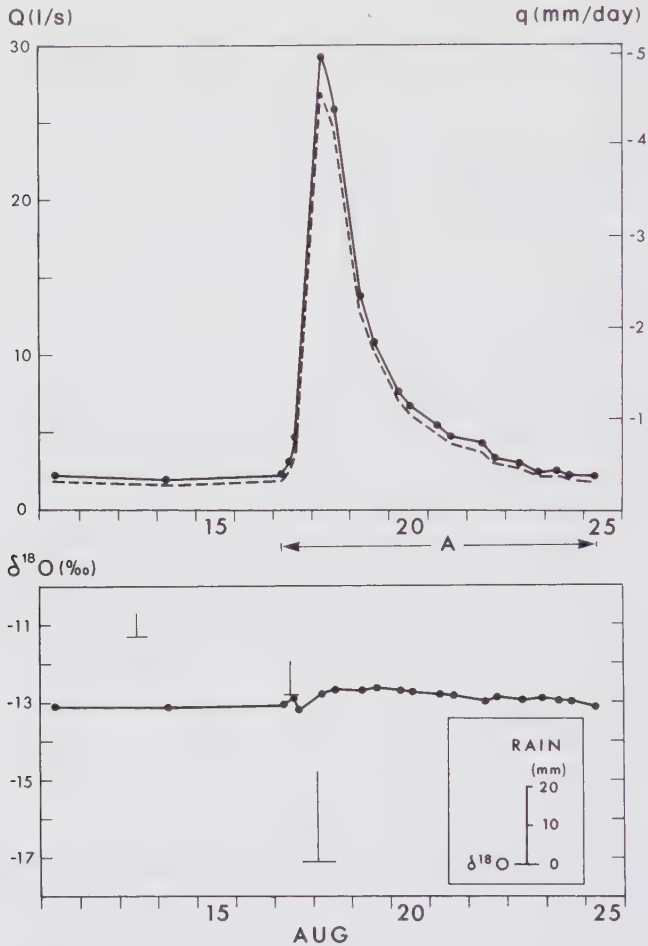


Fig.8.15 Stream hydrograph (unbroken line) with calculated groundwater hydrograph (broken line), $\delta^{18}\text{O}$ of streamwater, and amount and $\delta^{18}\text{O}$ of rainfalls. The length of the horizontal rainfall lines show the duration of rainfall when known; if the duration is not known the length marks the 24 h sampling period. Svartberget Nedre A, August 1981.

For Svartberget Nedre A, the existence of a substantial contribution of groundwater of high $\delta^{18}\text{O}$ from the mire (Svartberget Övre), might offer one further explanation for the unexpected variation in streamwater $\delta^{18}\text{O}$. There are no data for streamwater

$\delta^{18}\text{O}$ from Svartberget Övre from the period under consideration, but the data from the summer of 1982 show a significant difference in $\delta^{18}\text{O}$ between the two stations. The δ -values at the sub-basin Svartberget Övre were about 0.5‰ higher than those at Svartberget Nedre (Fig.A1.9).¹

Since the total flow at Svartberget Övre, during the time of peakflow at Svartberget Nedre, never exceeded 30% of the flow at the lower station, the maximum possible influence of a 0.5‰ difference in $\delta^{18}\text{O}$ is 0.2‰. This value is to be compared to the observed increase in streamwater $\delta^{18}\text{O}$ of about 0.5‰. Influence of the mire alone thus cannot explain the anomalous δ -values at Svartberget Nedre.

Hydrograph separation was performed using $\delta_g = -12.5‰$ (instead of initial streamwater = $-13.1‰$), assumed to represent the extreme case of groundwater contribution from Svartberget Övre only. The weighted mean of the two rainfall samples was used at these rough estimates. A groundwater fraction of 92% was obtained, this value being largely insensitive to errors in groundwater $\delta^{18}\text{O}$.

Even if the assumptions for these estimates are uncertain, it is concluded that the isotopic behaviour during the event can be explained only by a strong dominance of groundwater in streamflow.

For the other rainfall events, see Appendix 3.

8.2.3 Common features of the separated rainfall-generated runoff events

For only a few of the separated events could the groundwater fraction be determined with a sufficiently high time resolution to allow a detailed analysis of its variation over time (Gård-sjön Fl B, Stormyra A, Stormyra B). When separation could be performed during the hydrograph rise, these values usually showed the smallest groundwater fraction of the event. After

¹ The difference might be a consequence of a smaller groundwater reservoir at Svartberget Övre, causing less damping of the seasonal variation in precipitation $\delta^{18}\text{O}$ in this sub-basin, see Section 8.3.2. Enrichment during water storage in the mire might be another explanation to the difference observed in 1982.

peakflow the groundwater fractions tended to increase with time.

The rainwater and groundwater hydrographs from Stormyra A (Fig.8.12) probably show a general tendency of the time lag between the curves and their general appearances. The intensively sampled Gårdsjön F1 B and Stormyra B, as well as the less intensively sampled Nåsten A, Nåsten B, Stormyra C, and possibly also Svartberget Nedre C, conform, more or less distinctly, to this picture. The relative magnitudes of the two hydrographs may, on the other hand, differ greatly as shown by a comparison between Stormyra A and Gårdsjön F1 B.

The variation in groundwater fraction within the rainfall-generated events is similar to that found for snowmelt, but much more pronounced. The hysteresis effects of the curves of groundwater fraction versus total flow is thus much more striking for the rainfall-generated events. Since a possible modification by the channel net of the relationship between changes in tracer concentration and discharge would give a delay of the tracer change (Section 8.1.3), the observed advance of the tracer change must reflect a more rapid response of rainwater flow than of the groundwater flow to the stream.

Some of the investigated rainfall-generated hydrographs, Nåsten B, Nåsten E and Stormyra DI, show a small extra peak during the hydrograph rise. Similar peaks are sometimes met, upon rainfall or snowmelt, in these and other basins. Nice examples were found on days with large melting in Stormyra 1980. Such peaks probably originate from direct meltwater contribution, but there are no isotopic data available to confirm this statement. The results from Nåsten B suggest that it might be a correct hypothesis, but much denser sampling is needed for a safe conclusion.

8.2.4 Groundwater fractions of total discharged volumes

The total volume fraction of groundwater could be estimated for 14 events, see Table 8.6. The fractions by constant δ_g ranged between 68% and around 99% of the total flow, with a median value of 86%. For two further events, lower limits of the fraction could be estimated, the lowest being 53%. As discussed in Chapter 10, the volume fractions of groundwater show a weak tendency to decrease with increasing maximum runoff and also with total rainfall amount and rainfall intensity.

The sensitivity of some of the calculated groundwater fractions to systematical errors in δ_p of $\pm 1.0\%$ and in δ_g of $\pm 0.5\%$ are given in Table A3.1.

8.2.5 Extent of discharge areas

Basin fractions of discharge areas, calculated by Eq.(3.9), are listed in Table 8.6. The 14 values range between 0.2% and 17%, with a median value around 3%. The calculated fractions of discharge area are positively correlated with mean runoff during the rainfall and also with maximum runoff during the event (see further discussion in Chapter 10).

The only field survey carried out in connection with the rainfall-generated runoff events was performed in Aspåsen A. There the mapped discharge area was 15% of the basin, as compared with 7% obtained by isotopes (see Appendix 3). The largest fraction of discharge area found by isotopes among the events, 17% in Nåsten A, is not unrealistic for that basin at prolonged high flows in late autumn. The other high value, 15% met in Stormyra C, seems to be a little high for this summer event which was caused by a convective shower of a few hours' duration. It probably reflects the existence of Hortonian overland flow or of rainwater reaching the stream after having passed the soil (i.e., changing groundwater $\delta^{18}\text{O}$).

The small fractions of discharge area found at Svartberget Nedre, 0.3-3.2%, are not consistent with the physiography of this basin. Mires, which are expected to generate saturation overland flow, at least during late autumn, cover 20% of this basin. Even if a considerable part of the rainfall is stored in the mires, the difference between 20 and 0.3% seems high.

The fractions calculated for the other events in Gårdsjön, Nåsten, and Stormyra are considered realistic. At low groundwater levels the mires in these basins may contribute little to direct rainwater runoff. In Stormyra, a field survey at low flow shows that the drainage net constituted 0.2% of the basin area (Gotthardsson, 1973). This value should provide a lower limit for this basin of the fraction of discharge area as estimated by ^{18}O .

8.2.6 Conclusions on rainfall events basin by basin

Aspåsen

A medium event, occurring when the basin was very wet due to earlier rainfalls, was separated, showing that the flow was mainly groundwater. In contrast to the situation at the separated snowmelt event, the basin was almost completely clearcut at this time.

Buskbäcken

One very large summer event was investigated, but the isotopic conditions did not allow a quantitative hydrograph separation. The isotopic data suggest a clear dominance of groundwater during the event.

Gårdsjön F1 and F3

One medium and one small event were separated in each of these basins. Due to intense streamwater and rainwater sampling, the results, showing a strong dominance of groundwater in streamflow, are accurate. No significant difference was found between the basins.

Nåsten

The six events investigated cover the whole spectrum of peakflow magnitudes. Detailed pictures of the variations in groundwater fraction could not be obtained with the applied sampling-intensity, but the volume fractions could be accurately determined in most cases. Enrichment may have affected the results of two summer events (C and E), but even if this possibility is taken into account, the dominance of groundwater in streamflow is evident.

Stormyra

Complete separation was performed for three events of different magnitudes, and lower limits of groundwater fraction could be established for two additional events. The dominance of groundwater at peakflow in Stormyra C is noticeable. This very large

event was caused by a heavy thunderstorm with a rainfall intensity of around 50 mm h^{-1} . The isotopic data show a dominance of groundwater during all events.

Svartberget Nedre

Three events of large to medium size were investigated. Unexpected variations in streamwater $\delta^{18}\text{O}$ at one of the events, A, might be attributed to the mire in the upper part of the basin. Although the $\delta^{18}\text{O}$ of groundwater is uncertain for the events, possibly due to the mire, the large difference encountered between rainfall and initial streamwater $\delta^{18}\text{O}$ ensures that the dominance of groundwater found in the separations is real.

8.3 Estimates of reservoir volumes and transit times by ^{18}O

The isotope data presented for single runoff events in the preceding sections show that daily and in-storm fluctuations in the ^{18}O -content of precipitation are highly damped during the transformation of the precipitation into streamflow. Also, the seasonal variation in the ^{18}O -content of precipitation is damped to a large degree, as seen in the yearly summaries in Figs.A1.1-9. By the use of variable $\delta^{18}\text{O}$ of groundwater (Eq.3.6) in the hydrograph separations, the reservoir volume needed to simulate the damping from precipitation input to groundwater output during single events was estimated. In this section, reservoir volumes for water in the basins are estimated from the observed damping of the input variations over longer time periods, e.g., several months. The applied model is very simple, and since it could easily be developed to give a more refined analysis of the data, it is presented without details and just a few indicative results are shown.

By comparing the smoothed seasonal variations of $\delta^{18}\text{O}$ in the input (precipitation) with those of the output (streamflow), information on reservoir structures and volumes, as well as on mean transit times for water in basins has been obtained in several studies (e.g. Malozewski and Zuber, 1982, Burgman et al., 1987, cf. Section 2.3). In contrast to the approach in those studies, the model shown here is based upon daily values of precipitation $\delta^{18}\text{O}$, which are compared with daily (when available) values of streamwater $\delta^{18}\text{O}$. Thereby the large scatter in the $\delta^{18}\text{O}$ of the daily precipitation is utilized. A similar approach was used by Christophersen et al. (1984) and

Lindström and Rodhe (1986), who simulated the flow of ^{18}O through basins by already existing hydrochemical models, in order to calibrate the models by using a non-reactive tracer.

8.3.1 Short description of the model

The model structure is very simple. The basin is represented by a soil box which is directly connected with a mixing reservoir (Fig.8.16). The water flow through the soil box is calculated from water budget considerations for the root zone, in a similar way as in some runoff models (e.g. Bergström and Sandberg, 1983). Input variables are daily values of precipitation amount and $\delta^{18}\text{O}$, air temperature, and potential evapotranspiration. There is no time distribution function for the runoff, which occurs as pulses when field capacity is exceeded.

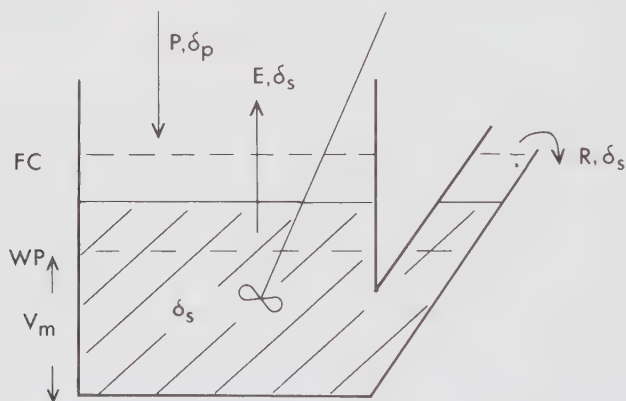


Fig.8.16 Model for estimates of reservoir volumes and transit times for water in a basin by using ^{18}O . P , E and R are precipitation (rainfall or snowmelt), evapotranspiration and runoff respectively, δ_p and δ_s the $\delta^{18}\text{O}$ of precipitation and soil water respectively, FC field capacity, WP wilting point and V_m the volume of mixing reservoir.

In calculating the flow of isotopes, complete mixing in the reservoir is assumed. Transpiration is presumed to be non-fractionating (cf. Section 6.2). The $\delta^{18}\text{O}$ of the transpiring water, as well as of the runoff, is then equal to the $\delta^{18}\text{O}$ of the soil water. Enrichment of the soil water by direct evaporation and molecular exchange with the atmosphere and enrichment during interception, both of which may be of some significance (cf. Section 6.2), are disregarded.

The isotope content of the snowpack is calculated from the $\delta^{18}\text{O}$ of the snowfalls. Neither enrichment of the snowpack by the atmosphere, nor fractionation between snow and meltwater is considered in this simplistic model. The net isotopic flux from the atmosphere (cf. Section 7.2.1) and a fractionation during melting can be introduced, however, to improve the calculations of meltwater $\delta^{18}\text{O}$ (cf. Lindström and Rodhe, 1986).

8.3.2 Results

The model was first calibrated to describe the water flow, mainly by varying the amount of plant available water. A close agreement between simulated and observed runoff is, of course, impossible to achieve with this simple model. However, when the agreement between simulated and observed accumulated runoff over longer periods was considered acceptable, the effect of various mixing volumes on the variations of the $\delta^{18}\text{O}$ of soil water and runoff was investigated.

In Fig.8.17 the $\delta^{18}\text{O}$ of soil water and runoff (which, in this model, are identical) calculated for various reservoir volumes are compared with observed $\delta^{18}\text{O}$ of streamwater. (For the input of ^{18}O by precipitation and the observed runoff, see Appendix 1, Figs.A1.3, A1.5 and A1.7.)

For the periods shown, the seasonal variations and also changes during many of the rainfall events, could be roughly simulated from precipitation data by choosing appropriate reservoir volumes. It is seen in the diagrams that the smallest volumes give too little damping of the input signals. A reservoir volume around 300 mm seems to describe the variations in Gårdsjön F1 and Nåsten and possibly a smaller volume, around 200 mm, in Stormyra. There is no significant difference in the isotopic behaviour of the streamwater in Gårdsjön F1 and F3 (cf. Fig.A1.4), both being comparatively well described by the model.

During winter and spring, there were large negative peaks in the $\delta^{18}\text{O}$ of streamwater in Nåsten. These variations could not be simulated by the model. The peaks were probably due to large fractions of meltwater during the initial phases of the snowmelt periods. (The spring floods in Nåsten and Stormyra 1980 could not be separated because of unfavourable isotopic situations, cf. Section 8.1.1.)

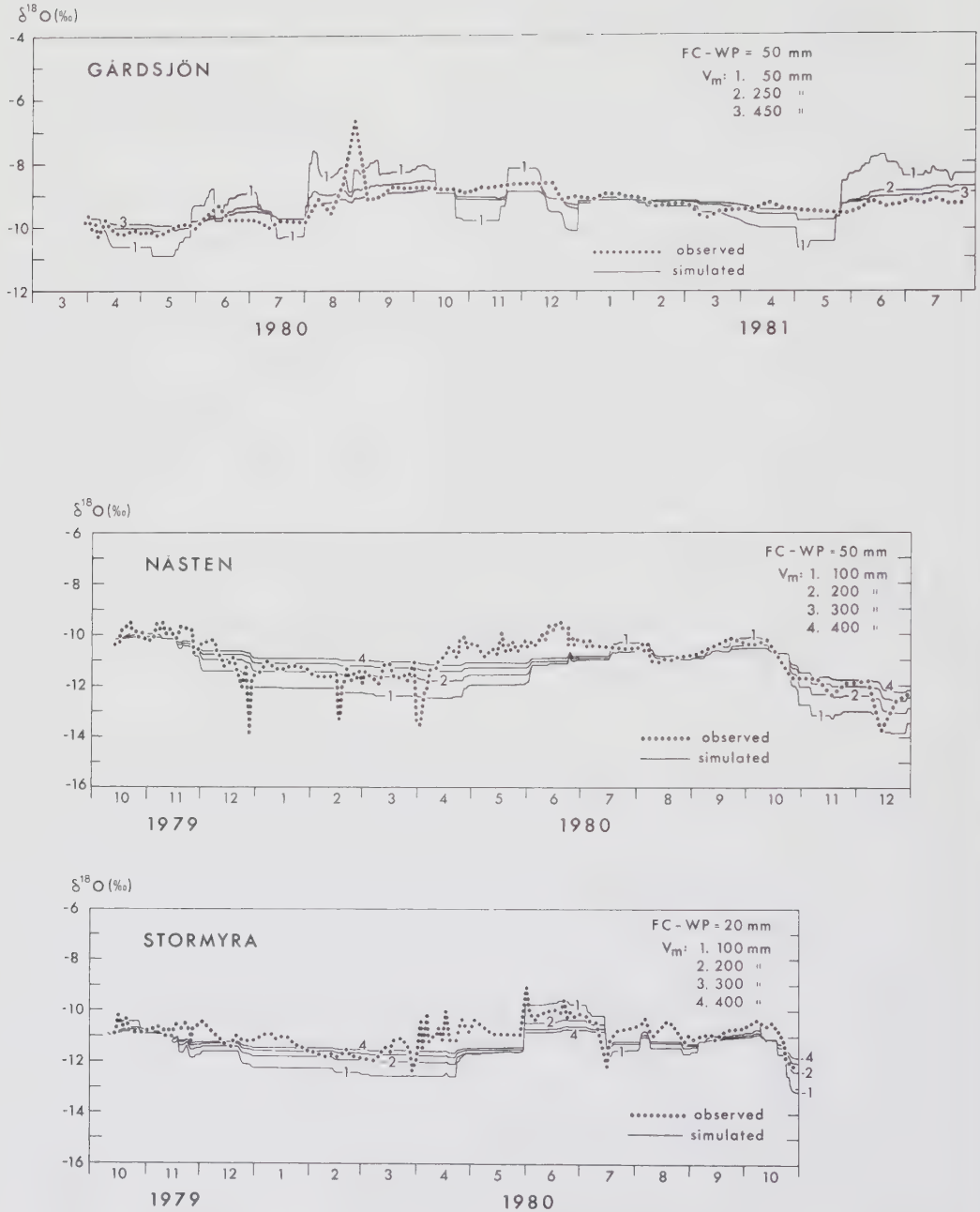


Fig.8.17 Observed and simulated $\delta^{18}\text{O}$ of streamwater. The simulations were performed with constant differences between field capacity (FC) and wilting point (WP) but with various volumes of the mixing reservoir (V_m). Gårdsjön, Nästen, and Stormyra.

In the above examples, the snowpacks were comparatively shallow and, consequently, the spring floods were modest. For periods with melting of deeper snowpacks, the model description of streamwater $\delta^{18}\text{O}$ was less successful. This result is not surprising, since no direct contribution of melt- or rainwater is included in the model. The hydrograph separations showed that the contribution of new water could be considerable during snowmelt, but it was generally small during rainfall events. For this reason, simulations were performed also for the snowfree periods in basins with large snowpacks.

If stationary flow is assumed, the mean transit time for water in the basins, t , can be calculated from

$$t = V/Q \quad (8.5)$$

where V is the reservoir volume and Q is observed long term mean runoff.

Table 8.7 Reservoir volumes and mean transit times estimated by ^{18}O .

Basin	Period	Reservoir volume (mm)	Mean transit time (years)
Buskbäcken	Summer periods 1981, 1982	300	0.7
Gårdsjön F1	March 1980 - July 1981	300	0.5
Gårdsjön F3	March 1980 - July 1981	300	0.5
Nåsten	Oct. 1979 - Dec. 1980 (June - Aug. 1982)	300 (50)	1.5
Stormyra	Oct. 1979 - Oct. 1980 (June - July 1982)	200 (<50)	0.7
Svartberget Övre	Aug. - Oct. 1981	200	0.6
	June - Oct. 1982	200	
Svartberget Nedre	Aug. - Oct. 1981	300	0.6
	June - Oct. 1982	300	

Reservoir volumes around 300 mm were obtained in most basins, giving mean transit times of 0.5 - 1.5 year (Table 8.7). The smaller reservoir volumes in Stormyra and Svartberget Övre may reflect different characteristics of these basins as compared to the others. In Stormyra the soil cover is very thin, and the

basin has large outcrop areas, giving little soil water storage. The smaller damping of the input signals in Svartberget Övre in comparison with Svartberget Nedre (cf. Fig.A1.9) indicates a smaller reservoir volume in Svartberget Övre. The difference is probably caused by the mire in Svartberget Övre, which allows little soil water storage. Enrichment of water due to evaporation from the mire in summer may, however, also contribute to the comparatively large seasonal variation in streamwater $\delta^{18}\text{O}$ at Svartberget Övre.

For summer periods of very little flow, there was a tendency for the $\delta^{18}\text{O}$ of streamwater to increase in a way that could not be simulated by the model, unless very small reservoir volumes were used (see values within brackets in Table 8.7). It seems probable that these increases were due to enrichment during stream channel flow (cf. Section 6.2.5). A similar increase in the $\delta^{18}\text{O}$ of streamwater, which could not be related to precipitation, was observed during summer low flow in the Birkenes basin in Norway (Christophersen et al., 1984).

In Chapter 10 the above reservoir volumes are compared with other estimates.

9 UNCERTAINTY DISCUSSION

The accuracy of the calculated groundwater fractions and fractions of discharge area depends first of all on the validity of the applied model. This question is discussed in other chapters. In this chapter the precision of the method is discussed.

As will be shown below, there are several possible systematical errors in the input variables. Due to the complexity of the systematical errors it is regarded as impossible, and statistically incorrect, to perform a thorough error analysis with the aim of providing statistical limits of confidence for the obtained results. For this reason, the analysis is performed as follows:

First, the sensitivity of the applied equations to errors in the input variables is investigated (Section 9.1). Next, various factors influencing the uncertainties in the input variables are identified. When possible, random or systematical errors due to single factors are quantified (Section 9.2). Finally, reasonable values of the uncertainties of the input variables are, rather arbitrarily, suggested and their influence on the results is shown (Section 9.3).

9.1 Sensitivity of the model to uncertainty in the input variables

9.1.1 Groundwater fractions

Groundwater fractions were calculated by Eq.(3.3),

$$X = \frac{c_s - c_p}{c_g - c_p}$$

The uncertainty in X , ΔX , due to random errors in the tracer concentration is obtained by differentiation as

$$|\Delta X| = \frac{|(c_g - c_p) \Delta(c_s - c_p)| + |(c_s - c_p) \Delta(c_g - c_p)|}{(c_g - c_p)^2} \quad (9.1)$$

or, using Eq.(3.3) once more to eliminate $c_s - c_p$,

$$|\Delta X| = \frac{|\Delta(c_s - c_p)| + |X \Delta(c_g - c_p)|}{c_g - c_p} \quad (9.2)$$

Assuming that Δc_s , Δc_p , and Δc_g are proportional to the standard deviations of their respective populations, and have the same constant of proportionality, we get by square summation

$$|\Delta X| = \frac{1}{c_g - c_p} \sqrt{\Delta c_s^2 + \Delta c_p^2 + X^2(\Delta c_g^2 + \Delta c_p^2)} \quad (9.3)$$

Thus, for a certain value of X , $|\Delta X|$ is inversely proportional to $c_g - c_p$. This difference can be said to define the isotopic condition for hydrograph separation of the event.

For a certain value of $c_g - c_p$, ΔX increases with X . But the relative uncertainty

$$\frac{\Delta X}{X} = \frac{1}{c_g - c_p} \sqrt{\frac{\Delta c_s^2 + \Delta c_p^2}{X^2} + \Delta c_g^2 + \Delta c_p^2} \quad (9.4)$$

decreases as X increases.

Eq.(9.3) gives the uncertainty for each calculation of X , i.e., for the instantaneous groundwater fractions. The total groundwater fraction during a period is calculated from n instantaneous values of X , at roughly constant time intervals of length Δt , by

$$X = \frac{\sum_{i=1}^n Q_i X_i \Delta t_i}{\sum_{i=1}^n Q_i \Delta t_i} \quad (9.5)$$

If the instantaneous values of ΔX are independent of each other, and if further Q is assumed constant and the error in Q is disregarded, the uncertainty in X is reduced to

$$X \approx \frac{\Delta X}{\sqrt{n}} \quad (9.6)$$

where n is the number of observations. But since the errors are not independent (cf. below), and since Q varies, giving the

largest contribution to ΔX from errors in X at the largest values of Q , the reduction will be less.

In calculating event mean values of X it was postulated that the instantaneous values of X , X_i , are close to the weighted mean of X during each time interval, Δt_i . During snowmelt, streamwater was predominantly sampled around the hours of maximum daily flow. The intensively sampled periods showed that in cases where there is a diel fluctuation in X , it is inversely related to that of discharge. Thus the applied sampling procedure tends to underestimate the period mean of X . From the intensively sampled periods this underestimation is estimated to be less than 5 percentage points.

9.1.2 Fractions of discharge area

Basin fractions of discharge area were estimated from event means of water input and meltwater flow by Eq.(3.8)

$$Y = \frac{Q}{A(P - \Delta S/\Delta t)} (1 - X)$$

Logarithmic differentiation gives the maximum error as

$$\left| \frac{\Delta Y}{Y} \right| \leq \left| \frac{\Delta Q}{Q} \right| + \left| \frac{\Delta A}{A} \right| + \left| \frac{\Delta(P - \Delta S/\Delta t)}{P - \Delta S/\Delta t} \right| + \left| \frac{\Delta X}{1 - X} \right| \quad (9.7)$$

Since the uncertainties in the input variables are probably independent, we get a more realistic relative accuracy in Y as

$$\frac{\Delta Y}{Y} \approx \sqrt{\left(\frac{\Delta Q}{Q}\right)^2 + \left(\frac{\Delta A}{A}\right)^2 + \left(\frac{\Delta(P - \Delta S/\Delta t)}{P - \Delta S/\Delta t}\right)^2 + \left(\frac{\Delta X}{1 - X}\right)^2} \quad (9.8)$$

9.2 Uncertainty in the input variables

9.2.1 Groundwater fractions

Factors affecting the uncertainty in δ -values used for calculation of X are listed in Table 9.1. When possible, a rough value is given of the uncertainty for which the various factors account. For systematical errors, the direction by which the error affects X is marked. Most of the factors are discussed in other sections. Three of the factors, not treated earlier, are discussed below; influence of precipitation during snowmelt, time lag between meltwater input and streamwater sampling, and

Table 9.1 Factors affecting the uncertainty in the input variables to Eq.(3.3) (fraction of groundwater)

	$\Delta\delta$ (‰)	Direction of influence ¹		
		Snow-	Rainfall	
		melt	$\delta_p < \delta_g$	$\delta_p > \delta_g$
All δ -values				
Analytical error	0.1			
$\delta^{18}\text{O}$ of streamwater, δ_s				
Enrichment during surface water flow (Section 6.2.5)	<0.5	+	+	-
$\delta^{18}\text{O}$ of precipitation, δ_p				
Snowmelt				
Areal variability of snowpack $\delta^{18}\text{O}$ (Fig.A2.1, Appendix 2)	1			
Isotope fractionation during melting (Sections 6.2.6, 7.2.1)	1	-		
Influence of vertical layering in the snowpacks (Section 7.1.2)	1			
Rainfall on wet areas (this section)		+		
Time lag between meltwater input and streamwater sampling (this section)	0.1	-		
Rainfall				
Temporal variation during the event (Section 7.1.2)				
$ \delta_p - \delta_g $ increases with time			-	-
decreases with time			+	+
Areal variability (Section 7.1.3)	<0.5			
Common				
Enrichment during overland flow (Section 6.2.5)		+	(+)	(-)
Molecular exchange with groundwater (this section)		+	+	+
$\delta^{18}\text{O}$ of groundwater, δ_g				
Influence of pre-event soil moisture (Section 8.1.2))	<1	-	$\pm$	$\pm$
Changing δ_g by infiltration (partly accounted for by using variable δ_g) (Section 8.1.2)	<1	-	-	-
Enrichment of streamwater before the event (only at low summer) (this section)	<0.5		-	+
Enrichment during saturation overland flow (Section 6.2.5)		+	+	-

¹ For systematical errors:

+ the error causes overestimation of the groundwater fraction
- the error causes underestimation of the groundwater fraction

molecular exchange between overland flow and groundwater. Moreover, some comments are made on the uncertainty due to enrichment during water flow.

Influence of precipitation during snowmelt

During snowmelt, δ_p is assumed to be equal to the basin mean of snowpack $\delta^{18}\text{O}$ at a certain moment. Possible modifying effects of precipitation are thus not accounted for in the calculations of X . These effects depend on whether the precipitation falls as snow or as rain and on whether it falls on snow-covered or on bare areas.

Snowfall on a snowpack will introduce little error, since changes of snowpack $\delta^{18}\text{O}$ are accounted for by the repeated snowpack samplings. Rainfall, with a δ -value differing from that of snowpack, may, on the other hand, affect δ_p . The largest influence will occur during large, intense rainfalls on small, melting snowpacks. The influence is difficult to quantify, since the rainwater is partly stored in the snowpack and also since there will be an isotopic exchange between rainwater and snow during the percolation through the snowpack. For these reasons, the ^{18}O supply by rainfall is partly included in the snow samplings and thereby partly regarded in the calculations of X .

Precipitation, of both types, on snowfree discharge areas may introduce errors in δ_p . If the snow falling on such areas melts completely before the next snow sampling occasion, snowfall and rainfall will have the same influence on δ_p , since they both contribute completely to streamflow.

It is to be noted that the uncertainty due to precipitation during snowmelt would not appear if δ_p was determined from representative snow lysimeters, since with such instruments the real water input to the ground is sampled.

Rough estimates of systematical errors on the groundwater fractions due to precipitation, according to a method given in Appendix 4, are given in Table 9.2. The smaller values, Δx_1 , are considered realistic, while the larger values, Δx_2 , probably represent upper limits of the errors. It should be emphasized that these estimates are very rough, performed with the aim of identifying periods with a possible large influence of precipitation. The time distribution of the precipitation amounts in

Table 9.2 Uncertainty due to precipitation during snowmelt.

Basin	Period		ΔX_1 (%)	ΔX_2 (%)
Aspåsen	1981	I	0	+2
		II	-3	-6
Buskbäcken	1980	I	-2	-6
		II	-1	-4
	1981	I		
		II		
		III		
Gårdsjön F1	1980		0	0
Gårdsjön F2	1980		0	0
Gårdsjön F3	1980		0	0
Nåsten	1979	I	-1	-1
		II	-1 ^a	-5 ^a
	1982	I	-4	-4
		II	-6	-16
		III	-2	-6
Stormyra	1979	I		
		II	-3 ^a	-5 ^a
	1982		-8	-20
Svartberget	1981	I	-5	-18 ^b
		II	0	0
	1982	I	+2	+2
		II	-5	-21

^a Incomplete data on precipitation $\delta^{18}\text{O}$

^b Very unlikely due to deep snowpack.

ΔX = correction to be added to the estimated total groundwater fractions to account for possible influence of precipitation on the $\delta^{18}\text{O}$ of the meltwater component.

ΔX_1 is calculated assuming direct contribution to streamflow by the precipitation falling on half the discharge area and

ΔX_2 assuming contribution from precipitation falling on an area equal to twice the discharge area (see Appendix 4).

relation to the discharge peaks is an important factor not regarded in these estimates.

Since the precipitation in spring usually is higher in ^{18}O than the snowpack, the groundwater fractions tend to be overestimated during periods with considerable precipitation. For most periods the estimated errors are small, resulting in an overestimation of the groundwater fractions of less than 5 percentage points, but for three of the periods, Nåsten II 1982, Stormyra 1982, and Svartberget II 1982, the error may not be negligible.

Time lag between meltwater input and streamwater sampling

When the groundwater fractions were calculated, meltwater $\delta^{18}\text{O}$ (or rather snowpack $\delta^{18}\text{O}$) on the day of streamwater sampling was used as δ_p . Daily values were obtained by interpolation between the days of snowpack sampling. Since the snowpack $\delta^{18}\text{O}$ gradually changes, and since it takes some time for the meltwater to reach the sampling point, an error may be introduced. Since snowpack $\delta^{18}\text{O}$ increases with time in almost all cases, and thereby also probably the $\delta^{18}\text{O}$ of meltwater, the error tends to underestimate the groundwater fractions.

Rough estimates using water velocities, stream lengths, flow, and reservoir volumes show that the time for meltwater to reach the basin outlets is unlikely to be more than a few days at most, and generally less than that.

Maximum error, for a time lag of two days, would be an overestimation of δ_p of around 0.4‰, found in Nåsten I 1979 (see Fig.A2.9). For most melting periods, the time lag would give an error in δ_p of around 0.1‰ or less. The resulting underestimations of groundwater fractions would be around 4% and 1%, respectively.

Molecular exchange between overland flow and groundwater

In the present study basins, and generally in Swedish till soils, saturated areas connected to the streams commonly occur during periods of high flow. With the hypothesis for streamflow generation sketched in Chapter 1 it is assumed that all new rain- or meltwater reaching the stream originates from rainfall or melting on such saturated areas, cf. Eq.(3.8). However, if overland flow is generated also outside the saturated areas (i.e. Hortonian overland flow) and this water passes the saturated areas on its way to the stream, the isotopic content of the overland flowing rain- or meltwater may change in direction towards that of the groundwater. In such a situation, parts of the "groundwater" reaching the stream according to the separation by ^{18}O is actually rain- or meltwater from outside the discharge area, and the groundwater fraction of streamflow would be overestimated. The situation can be compared with cold water flowing over warm ground, thereby increasing its temperature.

In Appendix 5 possible rates of isotopic change due to such a molecular exchange are estimated and related to the flow situations in the present study. The exchange can be of significance only if the upward particle velocity of the groundwater is less than the downward rate of tracer penetration by molecular diffusion. It is concluded that there could be an isotopic change by molecular exchange in some of the flow situations, with the largest effect during low-intensity snowmelt events. During high-intensity events, on the other hand, the effect can safely be neglected. Thus, the most striking result of this study, the comparatively large groundwater fractions observed also at the large events, is accurate. It is further pointed out that from a chemical point of view, errors in the hydrograph separations caused by isotopic exchange between the two flow components can be disregarded since the rainwater actually becomes groundwater by the exchange process, acting similarly on all constituents of the water.

Enrichment by the atmosphere

The effect of a possible enrichment in ^{18}O during water flow was not regarded in the calculations of groundwater fractions. In Section 6.2 attempts were made to estimate the magnitudes of possible enrichments at various stages of the water flow through a basin. Enrichment affecting hydrograph separation may occur in interception, overland flow, stream channel flow, and flow through reservoirs. The following conclusions apply to the hydrograph separations performed in the present study, but it should be emphasized that they are rather speculative, since there are no experimental data available except concerning interception.

Interception of rainwater may cause an enrichment of the precipitated water of a few tenths of per mil before it reaches the ground. The larger the intensity and the larger the total amount of rainfall, the less is the mean enrichment of the precipitated water. The effect is considered small for the separated events and has been disregarded.

Enrichment during overland flow could not be quantified. During snowmelt, the isotopic conditions favour enrichment of meltwater. But since the molecular exchange between meltwater on the ground and the atmosphere is of importance only in snowfree areas, the mean enrichment during melting of a basin-wide snowpack should be small. During rainfall events, little enrichment

of the precipitated water is expected during the rainfall, since the water is then probably close to isotopic equilibrium with a saturated atmosphere. After the rainfall, enrichment could take place, but then overland flow of rainwater rapidly ceases. Out-flowing groundwater could, on the other hand, become enriched during the rainfall and, particularly, after the cessation of rainfall.

There may be an enrichment during stream channel flow. Values of 0.5-1.0‰ were considered probable during low flow in summertime. At high flows, and in spring, the enrichment is smaller. The most important effect of this enrichment is probably to cause an overestimation of δ_g in some summer events with very small initial streamflow, since the $\delta^{18}\text{O}$ of stream-water before the event is assumed to be equal to δ_g . In spring the assessment of δ_g is not affected.

Temporary surface reservoirs develop during spring flood in some of the basins investigated. In some extreme cases, enrichment of 0.5-1.0‰ may occur. During most snowmelt events and all rainfall events, enrichment by flooding was small.

From the fact that the precipitated water may be either higher or lower in ^{18}O than the groundwater it follows that there are two basically different isotopic conditions for separation. Possible systematic errors due to enrichment would then act in opposite directions on the calculated groundwater fractions. During snowmelt events $\delta_p < \delta_g$ and all suggested enrichments tend to overestimate the groundwater fraction. During rainfall events with $\delta_p < \delta_g$ enrichment during the events similarly tends to overestimate the groundwater fractions. But here δ_g may be overestimated due to enrichment during stream channel flow before the event, tending to underestimate the groundwater fraction. In rainfall events with $\delta_p > \delta_g$ the effects will act in the opposite directions.

In Table 9.3 the rainfall events are divided into two groups, according to the isotopic situation, and the results of the separations are compared.

The finding that very similar results are obtained with both types of isotopic conditions indicates that the effect of enrichment is small during rainfall and thereby probably also during snowmelt.

Table 9.3 Calculated groundwater fractions for rainfall events having different isotopic conditions.

	$\delta_p - \delta_g$	
	<0	>0
Number of events	7	8
Mean groundwater fraction (%)	87	87
Median groundwater fraction (%)	81	86

9.2.2 Fractions of discharge area

Suggested relative uncertainties in the input variables in Eq.(3.8), by which the basin fraction of discharge area was calculated, are listed in Table 9.4. For systematical errors, the direction in which the error acts on the fraction of discharge area is given. Note the difference against Table 9.1 where the uncertainties were given absolutely, as δ -values (‰), whereas in Table 9.4 they are given as relative uncertainties.

When available, continuous water-level recordings have been used to calculate period mean discharge. The accuracy of the rating curve is assumed to be 5%. For some snowmelt events no water-level recordings were available, and period mean discharge was calculated from the water-levels at the moments of stream-water sampling. As pointed out in Section 8.2.1, this procedure may overestimate period mean discharge during snowmelt by up to 10%.

Eq.(3.8) gives the area needed, expressed as a fraction of the basin area, for the observed amount of rain or snowmelt to generate the rain- or meltwater flow estimated in the stream. Due to evaporation from discharge areas the real generation of rain- or meltwater flow on discharge areas exceeds the flow in the stream, $Q(1 - X)$. The error underestimates the fraction of discharge area by at most 10%, according to estimates for some summer events.

The relative uncertainty in the basin area generally increases as the area decreases, since the accuracy of the water divide is a certain number of metres rather than a fraction of the

Table 9.4 Uncertainty in the input variables in Eq.(3.8)

	Relative uncertainty (%)		Direction of influence + overestimate - underestimate	
	snow-melt	rain-fall	snow-melt	rain-fall
Period mean discharge, Q				
water-level recorder used	5	5		
water-level recorder not used	5-10		+	
evaporation from discharge areas		<10		-
Basin area, A	10	10		
Rate of water input, $P - \Delta S / \Delta t$				
uncertainty of measurements	5-50	10		-
interception				+
Fraction of new rain- or meltwater, $1 - X$	see the uncertainties in X estimated in next section			

basin area. 10% may be a high figure for the largest basins, but too small a figure for the smallest basins.

The precipitation measurements have not been corrected for aerodynamic and evaporative losses. They are moreover spot measurements, intended to represent the whole basins. During snowmelt, errors in P are disregarded, since P is generally small as compared to the rates of melting, determined from the change of snow water equivalent, ΔS . The uncertainty in S, i.e., in the water equivalent of the snowpack in the basin, is assumed to be 5 mm irrespective of S. The uncertainty in ΔS is then $\sqrt{5^2 + 5^2} = 7$ mm. With values of ΔS between 205 and 25 mm we get relative uncertainties in $P - \Delta S / \Delta t$ from 30 to 3 %, and choose a value of 10 %.

By rainfall interception the actual water input reaching the ground is smaller than the precipitation. The effect of interception, being relatively more important with small rainfall amounts, is counteracted by the measurement error of the precipitation amount. The two errors, both being around 10% for runoff-generating rainfalls, are assumed to balance each other.

9.3 Uncertainty of the calculated groundwater fractions and fractions of discharge area

We have now investigated the uncertainty in the variables used for calculations of groundwater fractions and basin fractions of discharge area. From these discussions, reasonable errors in the input variables are suggested and the resulting uncertainty of the results is shown, using the equations for the uncertainty of the applied models developed in the beginning of this chapter.

9.3.1 Groundwater fractions

The effect of systematic errors in δ_p due to precipitation during snowmelt was estimated separately for each separated event (Table 9.2). Systematic errors in δ_g , due to changes by infiltration, are partly regarded in the calculations using variable δ_g . If the groundwater fractions are interpreted as "fractions of pre-event water", these systematic errors in δ_g disappear.

The total effect of the other factors listed in Table 9.1 on the uncertainty of the input variables in Eq.(3.3) is suggested according to Table 9.5.

Table 9.5 Suggested uncertainty in the input variables in Eq.(3.3)

	Snowmelt	Rainfall
$\Delta\delta_s$	0.2	0.2
$\Delta\delta_p$	1.0	0.5
$\Delta\delta_g$	0.3	0.3

The uncertainty of the calculated groundwater fraction is given by Eq.(9.3), shown graphically in Fig.9.1 for different values of $\delta_g - \delta_p$, assuming the errors in the input variables from Table 9.5. The figure shows the uncertainty in single values of X caused by random errors in the input variables. Median values of $|\delta_g - \delta_p|$ in the snowmelt events were 3.9‰ at the beginning of melting and 2.7‰ later in spring, giving uncertainties around 0.3 to 0.5 for single values of X . In the rainfall events the median value was 3.7‰ and the uncertainty according to Fig.9.1 around 0.2. For period mean values of X , the uncertainty

may decline to $\Delta X/\sqrt{n}$ (Eq.9.6). Typical values of n are 15 for a snowmelt event and 8 for a rainfall event, reducing the uncertainty to 0.08 - 0.13 for the snowmelt events and to 0.07 for the rainfall events. As pointed out in Section 9.1.1, Eq.(9.6) yields a lower limit to the uncertainty. The actual value is not considered possible to quantify.

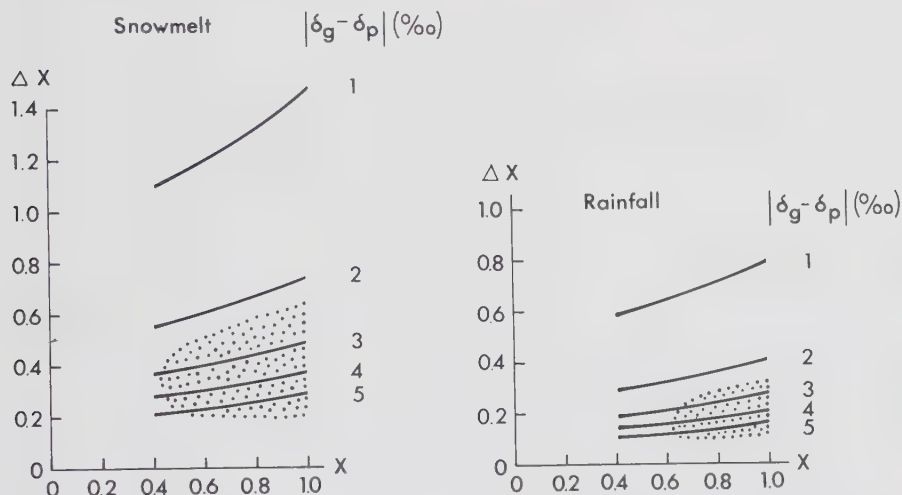


Fig.9.1 Uncertainty (ΔX) in single calculations of groundwater fractions (X) at the assumed uncertainties in the input variables (Table 9.5). The uncertainty is given for various differences between the $\delta^{18}\text{O}$ of the groundwater component (δ_g) and of the rain- or melt-water component (δ_p). Dotted areas symbolize typical combinations of variables found in the events investigated.

The sensitivity of the calculated period mean groundwater fractions during snowmelt was also investigated by calculating these fractions assuming systematical errors in δ_p of $\pm 1.0\text{‰}$ and in δ_g of $\pm 0.5\text{‰}$. The total effect of the systematical errors was then calculated as $\Delta X_{\text{tot}} = \sqrt{\Delta X_p^2 + \Delta X_g^2}$, where ΔX_p and ΔX_g are the sensitivity to the error in δ_p and δ_g respectively (systematical errors in δ_s are disregarded), see Table A2.1. For the respective spring flood totals, ΔX_{tot} ranged from ± 0.10 (Gårdsjön F3 1980) to about ± 0.3 (Aspåsen 1981), median value ± 0.19 . Median values of $\Delta \delta_p$ and $\Delta \delta_g$ was ± 0.16 and ± 0.09 respectively.

9.3.2 Fractions of discharge area

Uncertainties, estimated by Eq.(9.8), of the basin fractions of discharge area according to random errors are given in Table 9.6. The uncertainty of the total volume fraction of groundwater

is assumed to be 0.10 for snowmelt and 0.05 for rainfall events, and the uncertainties of the other input variables in Eq.(3.8) are assumed to be according to Table 9.4.

As for the groundwater fractions, the total sensitivities of the calculated basin fractions of discharge area to systematical errors in δ_p and δ_g are given for snowmelt periods in Table A2.1.

Table 9.6 Relative uncertainties of basin fractions of discharge area for various fractions of groundwater.

Type of event		Mean fraction of groundwater, X				
		0.5	0.6	0.7	0.8	0.9
Snowmelt	$\Delta S/S = 0.05$	0.23	0.28	0.36	0.51	1.01
	$\Delta S/S = 0.50$	0.55	0.57	0.67	0.72	1.12
Rainfall		0.25	0.29	0.37	0.52	1.01

$\Delta S/S$ = relative uncertainty of the change in snow storage

9.4 Conclusions on uncertainty

The uncertainty in the calculated groundwater fractions is inversely related to the difference between the $\delta^{18}\text{O}$ of groundwater and rain- or meltwater, which difference defines the isotopic condition for hydrograph separation of the event. With the suggested random uncertainties in the input variables in Eq.(3.3), the uncertainty in single estimates of the groundwater fraction may be considerable (Fig.9.1). But looking at a whole event, the effect of random errors is greatly reduced. The uncertainty of the figures given on period mean groundwater fractions may be around ± 0.10 for snowmelt events and somewhat smaller for most of the rainfall events.

A large part of the uncertainty in the groundwater fractions, particularly during snowmelt, is caused by the uncertainty in the $\delta^{18}\text{O}$ of the rain- or meltwater component. For snowmelt events, the accuracy would be considerably improved if δ_p were measured directly, by the use of representative snow lysimeters. However, variations within the events of the $\delta^{18}\text{O}$ of the water leaving the snowpack, and of the $\delta^{18}\text{O}$ of rainfall, may prevent accurate estimates of the time variation of the groundwater fraction, even if the δ -values were accurately determined. A more elaborate model for hydrograph separation is needed if

short-term variations in the $\delta^{18}\text{O}$ of the water input are to be accounted for.

The relative uncertainty in the figures given on basin fraction of discharge area is large, being for many snowmelt events around ± 0.5 .

Several systematical errors have been identified. Estimates have shown that, except for the error caused by rainfall during some of the snowmelt events and the underestimation of the groundwater fraction due to changes in the $\delta^{18}\text{O}$ of the groundwater component, the general results obtained in the study are not biased by systematical errors.

10 CONCLUSION AND DISCUSSION ON STREAMFLOW GENERATION

In the first part of this chapter, the results of the hydrograph separations are summarized. The estimated fractions of groundwater and basin fractions of discharge area are related to the hydrological conditions during the events. Reservoir volumes for water in the basins, obtained by different methods, are commented upon, as are the main methodological problems encountered in interpreting streamwater $\delta^{18}\text{O}$ in terms of runoff processes in the basins.

In the second part of the chapter, the process of streamflow generation is discussed on the basis of the results obtained from this study and a hypothesis is presented. The discussion is held in general terms, applying basic groundwater flow relationships to idealized hillslopes. The aim is to identify and evaluate various mechanisms by which rainwater or meltwater is transformed into streamwater.

10.1 Conclusions

10.1.1 Total volume fractions of groundwater

The hydrograph separations by means of ^{18}O show that storm runoff from the investigated basins, situated in different parts of Sweden, is usually dominated by groundwater (Table 10.1, see individual results in Tables 8.1 and 8.6). The dominance is univocal for all rainfall events and most of the snowmelt events.

Table 10.1 Summarized results of hydrograph separations by ^{18}O

	Fraction of groundwater by constant $\delta^{18}\text{O}$ of groundwater (%)				Fraction of discharge area (%)			
	n	min	median	max	n	min	median	max
Snowmelt events	29	32	61	95	22	2	22	60
Spring flood totals	13	42	59	86				
Rainfall events	14	68	86	99	14	0.2	3	17
All events	43	32	79	99				

n = number of events

As was discussed in other sections, the groundwater fractions estimated by ^{18}O rather refer to fractions of pre-event water, i.e., water that existed as groundwater or soil water in the basin before the events. The actual fractions of groundwater are larger, since some of the new rain- or meltwater observed in the stream reached the stream as groundwater after having infiltrated into the soil during the event. Attempts were made to estimate the change of tracer concentration of the groundwater component, resulting in an increased estimate of the groundwater fraction of about 10 percentage points or less. However, these fractions, too, are considered underestimates.

In order to generalize the results, the total volume fractions of groundwater estimated by ^{18}O using constant groundwater $\delta^{18}\text{O}$ are related below to hydrological conditions during the events. Since there are only a few events investigated in each basin, data from all basins have been put together to form one population. When applying this procedure it is presumed that the volume fractions of groundwater depend on general flow conditions rather than on local geological, geomorphological, or climatic conditions.

The closest relationship between groundwater fraction and a single hydrologic variable was the one found with maximum specific discharge of the event (coefficient of correlation $r = -0.68$ for the whole group of events, Fig.10.1 and Table 10.1). Fig.10.1 shows a systematic difference between rainfall and snowmelt events, in that the rainfall events have some 10 - 20-percentage-point higher groundwater fractions than the snowmelt events for a certain specific discharge.

Smaller coefficients of correlation were found with water input to the basins. Fig.10.2, showing groundwater fraction versus mean rate of water input for the snowmelt events, is discussed later in this chapter. For the rainfall events, maximum rainfall intensity gave about the same coefficient of correlation as did total rainfall amount, about -0.5 , both coefficients being smaller than the one with maximum specific discharge (-0.68 for the rainfall events).

The comparisons so far concern fractions of groundwater by constant $\delta^{18}\text{O}$ of groundwater, i.e., fractions of pre-event water. The finding that the fraction of pre-event water decreases as the flow increases seems natural, since the fraction of newly infiltrated water in the groundwater discharge then also increases. But the negative correlation remains for groundwater

fractions calculated by variable $\delta^{18}\text{O}$ of groundwater ($r = -0.70$ for such groundwater fractions versus maximum spring flood specific discharge). This relationship is not self-evident, but it is in accordance with the hypothesis for streamflow generation presented schematically in Chapter 1 and discussed in detail in later sections of this chapter.

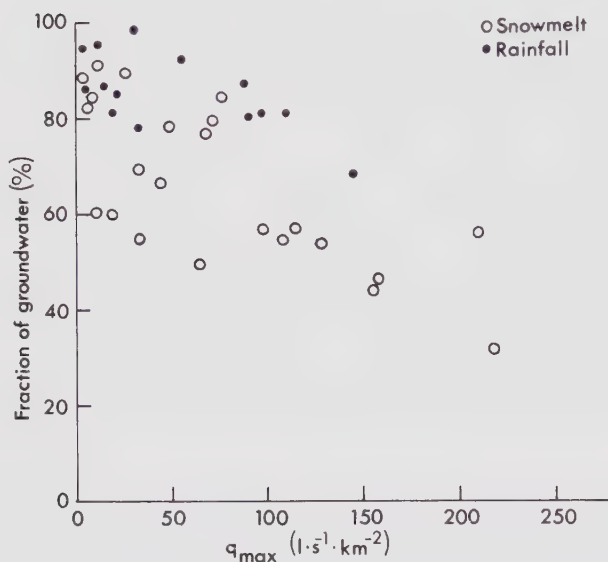


Fig.10.1 Total volume fraction of groundwater versus maximum streamflow. Snowmelt and rainfall events.

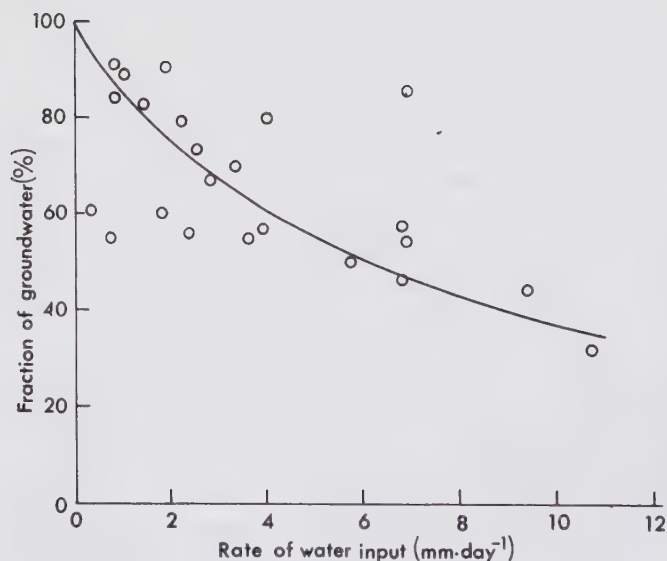


Fig.10.2 Total volume fraction of groundwater versus mean rate of water input. Snowmelt events.

Table 10.2 Linear regression between total volume fraction of groundwater and basin fraction of discharge area, as estimated by ^{18}O , and hydrological conditions during the events.

Dependent variable	Group of events	Independent variable	Number of events	Coefficient of correlation
Fraction of groundwater (constant $\delta^{18}\text{O}$ of groundwater)	Snowmelt events	$q_{\max}$	23	-0.73
		$P - \Delta S / \Delta t$	23	-0.65
	Rainfall events	$q_{\max}$	14	-0.65
		$i_{\max}$	11	-0.51
		$P \Delta t$	14	-0.46
	All events	$q_{\max}$	37	-0.68
Fraction of discharge area	Snowmelt events	$q_{\max}$	22	0.83
		$\bar{q}$	22	0.78
		$P - \Delta S / \Delta t$	22	0.56
	Rainfall events	q_{rainfall}	14	0.79
		$\bar{q}$	14	0.79
		$q_{\max}$	14	0.72
		$X \cdot \bar{q}$	14	0.73
	All events	$\bar{q}$	36	0.80
		$q_{\max}$	36	0.75
		X	36	-0.83
		$P \Delta t - \Delta S$	36	0.55
		$X \cdot \bar{q}$	36	0.60

$q_{\max}$ = maximum specific discharge

$\bar{q}$ = mean specific discharge

q_{rainfall} = mean specific discharge during the time of rainfall

P = rate of precipitation ($P \Delta t$ = total rainfall amount)

$\Delta S / \Delta t$ = change of snow storage over the time Δt

$i_{\max}$ = maximum 15-min rainfall intensity

X = total volume fraction of groundwater ($X \bar{q}$ = mean groundwater flow)

10.1.2 Extent of discharge areas

The areal extent of discharge areas was estimated from the hydrograph separations assuming that all rainfall and meltwater on recharge areas infiltrates, whereas all such water on discharge areas generates overland flow and is discharged in the stream during the event (Table 10.1, see individual results in Tables 8.1 and 8.6). Field surveys during some of the snowmelt events and one rainfall event showed that the extent of the discharge areas is reasonably well estimated (Fig.8.8).

For a single basin, the instantaneous fraction of discharge area is expected to increase with groundwater level and thus with groundwater flow. But since only mean values for the periods of water input could be calculated here, such comparisons could not be made. As with the volume fractions of groundwater, the estimated basin fractions of discharge area from all basins are put together to form a population large enough for meaningful comparisons with the flow situations.

Fig.10.3 compares calculated mean fractions of discharge area with period means of calculated groundwater flow. There is a positive linear correlation ($r = 0.60$, Table 10.1), but the points are very scattered. For a certain mean rate of groundwater flow, the calculated fraction of discharge area is generally larger at snowmelt events. One reason for this difference is that Eq.(3.9) gives the fraction of discharge area during the actual time of water input. For the rainfall-generated events, this time is short as compared to the duration of the flow event. During snowmelt, on the other hand, the water input occurs over a large fraction of the flow duration. The groundwater flow during the time of water input was generally about half of the period mean value for the rainfall events, but about equal to the period mean for the snowmelt events.

It is to be noted that the fraction of discharge area, Y , and the groundwater flow, Q_g , are both calculated from the groundwater fraction, X , and the total streamflow, Q . For this reason, the two variables are a priori related to each other. But since Y is proportional to $(1 - X)Q$ (Eq.3.9), while $Q_g = X Q$, a positive correlation would not be fictitious, provided that there are sufficiently large relative variations in X , which is the case here. As discussed later, the relationship follows from the relationship found between groundwater fraction and rate of water input.

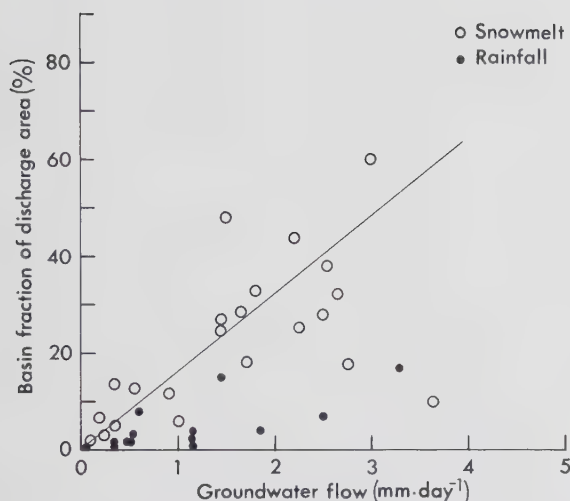


Fig.10.3 Basin fraction of discharge area, as estimated by ^{18}O , versus period mean of groundwater flow. Snowmelt and rainfall events.

A straight line of Y versus Q_g in Fig.10.3 would indicate that the groundwater discharge per unit discharge area is constant, equal to the slope of the line. The line drawn in Fig.10.3, for instance, corresponds to a groundwater discharge per discharge area of about 6 mm/day. With more data available, the following analysis might be made on data from single basins: A positive intercept with the Y -axis would give the extension of permanent discharge areas, which are saturated also during periods of very little flow. A positive intercept with the Q_g -axis, on the other hand, would show that there is a considerable groundwater discharge directly to the stream.

Since the groundwater hydrographs roughly follow the total hydrographs, due to small relative variations of the groundwater fractions within the events as compared to those in the total flow, the fraction of discharge area is also positively correlated to total flow (Table 10.2). The observed high coefficient of correlation ($r = 0.80$) does not, however, imply the existence of a physical relationship. This is so since Y is by definition (Eq.3.9) positively correlated with Q due to larger relative variations in Q than in X . The relationship (not shown here) might, however, be of more practical interest than the one shown in Fig.10.3, since it relates the basin fraction of discharge area to a directly measurable variable, mean total discharge (cf. the good relationship observed by Myrabö, 1986, from field surveys in a small Norwegian basin).

10.1.3 Reservoir volumes

Information about reservoir volumes for water in the basins, i.e. about the volume of soil water and groundwater participating in the water flow to the streams, has been obtained in the study in three different ways:

1. The total volumes of pre-event water discharged during the events, i.e., the volumes of groundwater calculated by constant $\delta^{18}\text{O}$ of groundwater, establish the lower limits of the amounts of water stored as soil water or groundwater before the events. These volumes are obtained directly from Tables 8.1 and 8.6 as $R \cdot X$, where R is total runoff and X the groundwater fraction.

For the respective spring-floods totals, the volumes of pre-event water ranged from 15 mm (Gårdsjön F1 1980) to 93 mm (Stormyra 1979), with a median value of 62 mm. For the rainfall events they ranged from 1 to 33 mm, median value 7 mm. These figures are certainly much smaller than the actual reservoir volumes, since far from all soil water and groundwater was replaced by new water during the events. This is seen from the fact that the $\delta^{18}\text{O}$ of streamwater after the events, when there is only groundwater flow to the streams, still differed considerably from that of rainwater. Different results were obtained in preliminary hydrograph separation by ^{18}O in the Birkenes basin in Norway (Christophersen et al., 1984). In this basin, the reservoir content seemed to have been replaced by new water during a large snowmelt event with considerable rainfall (≈ 130 mm discharged pre-event water).

2. Information on reservoir volumes for water was also obtained when using variable $\delta^{18}\text{O}$ of groundwater (Eg.3.6) in the hydrograph separations. Those estimates yielded the constant reservoir volume needed to generate the observed net changes of groundwater $\delta^{18}\text{O}$ during the events, under the assumption of complete mixing with the infiltrating water. As discussed in Section 8.1.3, this procedure applied to the spring floods yielded reservoir volumes of 100 - 300 mm. As with the figures given under point 1 above, these figures include both soil water and groundwater.

3. The most realistic reservoir volumes are probably those obtained by the model presented in Section 8.3. With this model, attempts were made to simulate the variation of streamwater $\delta^{18}\text{O}$ over periods of several months, using daily amounts and

$\delta^{18}\text{O}$ -values of precipitation. It was then assumed that all streamwater originated from soil water or groundwater. The completely mixed soil water reservoir, which was connected to a mixing reservoir, was allowed to vary according to the calculated water budget of the root zone. With the exception of large snowmelt events and some summer periods of very low flow, the $\delta^{18}\text{O}$ of streamwater could be roughly described by the model, using reservoir volumes around 200 - 300 mm (Table 8.7). With these reservoir volumes, mean transit times for water of around 0.5-1.5 year were obtained.

^{18}O data from two of the basins have also been used in the more elaborate HBV-PULSE model, which gave considerable improvement in reproducing the observed ^{18}O -values of the streams (Lindström and Rodhe, 1986). With this model, reservoir volumes around 480 mm and 440 mm were obtained for Buskbäcken and Gårdsjön F1 respectively, as compared with about 300 mm in both basins with the present simple model. The larger volumes obtained by Lindström and Rodhe may partly be due to the assumption of a bypass in their model, by which a certain fraction of the streamflow is fed directly by new rain- or meltwater.

There are not sufficient data available for detailed estimates of the soil water and groundwater storage in the basins from the thickness of the soil cover and the observed water content of the soil. Assuming a mean volumetric water content of the soil of 30%, water storages of 200 and 400 mm correspond to a soil thickness of 0.7 and 1.3 m respectively. The figures agree qualitatively with the figures on mean till depth in the Gårdsjön basins, around 2.1-2.6 m (Olsson et al., 1985), and on mean soil depth in the Masbybäcken basin, close to the Buskbäcken basin, of 1.8 m (Lundin, 1982). The water storage in Svartberget, on the other hand, is probably larger than a few hundred millimetres, since the soil cover in this basin may be several metres in thickness.

10.1.4 Methodological problems

Several methodological problems in hydrograph separation by ^{18}O have been discussed throughout the text and in the uncertainty discussion. These discussions mainly focus on the precision of the method, i.e., on the uncertainties in the input variables and the sensitivity of the calculated groundwater fractions to these uncertainties.

A severe shortcoming in the available data is the lack of systematical measurements of the isotopic composition of meltwater. The assumption that meltwater $\delta^{18}\text{O}$ is equal to the basin mean $\delta^{18}\text{O}$ of the snowpack at any moment introduces a comparatively large uncertainty in the groundwater fractions calculated for snowmelt periods and prevents a detailed analysis of the meltwater and groundwater hydrographs.

Due to the complexity of the uncertainties involved, it was not considered realistic to give statistical limits of confidence for the calculated groundwater fractions. With the suggested uncertainties in the input variables, the accuracy of the total groundwater fractions mostly fall around ± 15 percentage-points.

Two possible systematic errors, which have not been commented upon in the literature, have been discussed in some detail: isotope enrichment of the off-running water by molecular exchange with the atmosphere, and isotope changes of overland flowing groundwater due to molecular exchange with groundwater in saturated areas. Theoretical considerations show that these errors cannot confidently be ignored for all events, but due to lack of experimental data the figures given on the effect of these processes are tentative. The effect of enrichment would be largest in snowmelt events with considerable flooding, and the effect of molecular exchange in events with low intensity of water input per unit of saturated area without groundwater outflow, i.e., in low intensity snowmelt events in basins in which the stream is surrounded by large flat areas. In both of these cases, the error gives an overestimation of the groundwater fraction. In moderate- to high-intensity rainfall events without flooding, on the other hand, the errors can be ignored.

Two of the rainfall events gave unexpected variations of streamwater $\delta^{18}\text{O}$, showing that streamwater was not a mixture between rainwater with the observed $\delta^{18}\text{O}$ of rainfall and groundwater with a δ -value equal to that of streamwater before the event (Figs. 8.14 and 8.15). In both cases the $\delta^{18}\text{O}$ of streamwater increased, although the $\delta^{18}\text{O}$ of rainfall was considerably lower than that of streamwater before the event. Transformation of pre-event soil water of high $\delta^{18}\text{O}$ into groundwater during the events was suggested as an explanation for the anomalous behaviour. High δ -values of pre-event soil water for these events are expected from rainfall data. Enrichment of soil water by molecular exchange with the atmosphere is another, less probable, explanation. Similar anomalies were observed in two events in the Birkenes basin (Christophersen et al., 1984).

These events were preceded, however, by rainfalls of low δ -values. The existence of "anomalous events" points towards a shortcoming in the method of hydrograph separation by isotopes in its present application. Detailed investigations of such events may, however, throw new light on the process of stream-flow generation.

10.2 Discussion

Hydrograph separation by means of ^{18}O relies on measurements at mainly one point in a basin, the outlet. It gives information on the integrated result of the processes by which water is delivered to the stream, but little or no information is obtained on the various processes involved. The aim of the following sections is to discuss possible mechanisms for stream-flow generation which can explain the results obtained by ^{18}O . With this starting point, the discussion is necessarily limited to a general level and the suggested models are very rough.

The more important isotope studies reported in the literature were presented in Section 2.3. Those study basins are all within the temperate climate zone, but physiographic and geologic factors vary largely, from steep Alpine catchments to agricultural land of low relief. There are also large differences in the rates of water flow in the investigated events. Nevertheless, the results are similar in practically all these studies. The present study, and the isotope studies reported in the literature, all indicate that runoff events in small streams in temperate areas are normally dominated by pre-event water. Even if exceptions do occur, the phenomenon seems to be general. From the available results it has not been possible to identify physiographic factors which govern the role of pre-event water in storm runoff.

Considering the above results, the following questions arise:

1. By what mechanism is the input impulse from precipitation transformed into increased outflow from the soil to the stream?
2. Where in the basins are the impulses for storm runoff given?
3. From where in the basin do the off-running water particles originate?

4. For how long a time have the discharging water particles stayed in the basins, i.e., what is the transit time distribution of the discharging water?

The answers to these questions certainly differ when different basins are studied. The following discussion refers to the conditions in Swedish till soils.

10.2.1 Hypothesis for streamflow generation in till soils

The infiltration capacity of the soil is usually larger than the intensity of rainfall or snowmelt (cf. Section 2.1.1). Thus all water on unsaturated areas infiltrates. The water table largely follows the topography, with a depth below ground surface of a few metres in elevated areas. In the lower parts of certain hillslopes and in some low-lying areas, the water table is at or above the ground surface. These saturated areas generally, but not always, occur in connection with the streams. Depending on whether the water flows into or out of the groundwater zone, the area of a basin can be divided into recharge or discharge areas for groundwater. The areal extent of discharge areas is highly variable, expanding with rising groundwater level. During dry periods the discharge area may constitute only the stream itself.

The outflow of groundwater in discharge areas may respond rapidly to rainfall or snowmelt on recharge areas, causing a substantial groundwater component in stream stormflow. This groundwater reaches the stream either as overland flow on saturated areas or directly through the stream bottom and sides. All rainfall on discharge areas (except water lost by evapotranspiration, particularly during interception) reaches the stream as overland flow. Together with the groundwater discharged outside the stream it forms the saturation overland flow.

Discharge areas can be defined as "areas in which the uppermost groundwater has a flow component directed away from the groundwater zone". The flow direction is compared with the plane of the water table and not with the horizontal plane. But since the slope of the water table generally is small as compared with possible flow directions, a sufficiently accurate definition mostly is "areas in which the groundwater has an upward flow component". In the corresponding way recharge areas can be

defined as "areas in which the groundwater has a downward flow component".

With the above definition, a discharge area may be saturated or unsaturated, depending on whether the groundwater table reaches the ground surface or not (Grip and Rodhe, 1985). In a saturated discharge area no infiltration can take place and all rainfall generates saturation overland flow together with the outflowing groundwater. In an unsaturated discharge area, the upflowing groundwater is either lost by evapotranspiration from the unsaturated zone or linked off laterally through layers of high hydraulic conductivity. During wet periods there may be a downward water flow in the unsaturated zone and at the same time an upward flow in the saturated zone. A prerequisite for this lateral linking-off of the upflowing groundwater is that the saturated hydraulic conductivity increases towards the ground surface. Unsaturated discharge areas are probably common in till soils. The ground surface may therefore remain unsaturated also during wet periods at locations where one, from a topographic point of view, would expect groundwater discharge.

10.2.2 Limitations of the estimates of discharge areas by ^{18}O

With the above definitions, the model sketched in Fig.3.1 needs some modification. The discharge areas estimated from the hydrograph separations by Eq.(3.9), and also those determined by field surveys, constitute the saturated discharge areas. Eq.(3.9) gives the area needed to collect the volume of rainwater or meltwater discharged by the stream. But if the unsaturated zone is very thin and further if it consists of coarse material, which is the case in areas where the upflowing groundwater is linked off laterally, the capacity for storing water in the zone is limited. Thus, if the particle velocity of the lateral groundwater flow is high enough, the unsaturated discharge area may also contribute to the collection of the new rain- and meltwater that is discharged in the stream during an event.

Eq.(3.9) should be treated as a crude attempt to interpret the hydrograph separations by means of ^{18}O in terms of processes within the basin. For this reason, the distinction between saturated and unsaturated discharge areas was not made in the analyses of the hydrograph separations. Neither are the two types of discharge areas distinguished in the discussion on streamflow generation that follows.

When estimating the basin fraction of discharge area by Eq.(3.9) it is further assumed that the only saturated areas in a basin are the discharge areas (or rather the saturated discharge areas). One necessary condition for this equivalence is, of course, that no surface saturation by rainfall excess occurs. But even if this condition is fulfilled, the ground may be saturated also in areas where the groundwater flow has a downward component, i.e., in recharge areas. The situation could occur during extremely wet periods when the water table is at or above the ground surface along the main part of the hillslopes. Then there would be a gradually increasing infiltration with increasing distance above the point of transition from discharge to recharge, the so-called hinge point (Freeze and Cherry, 1979), as shown in laboratory experiments by Abdul and Gillham (1984). However, as long as Eq.(3.9) gives basin fractions of discharge area of less than 50%, which it does here with one exception, saturated recharge areas are probably of little importance.

10.2.3 Time variation of the groundwater fraction and the basin fraction of discharge area.

From the hypothesis for streamflow generation it follows that the new rainwater flow in the stream, at a certain rainfall rate, should increase with increased groundwater level, since the extent of overland flow producing discharge areas then increases. The hydrograph separations by ^{18}O have shown that the groundwater flow, with minor deviations, generally varies in parallel with the total flow. Thus the groundwater level, and, consequently, the extent of discharge areas, increases with increasing total flow (Fig.10.3, Table 10.2). Increasing discharge areas does not, however, directly imply that the groundwater fraction of streamflow decreases; this is a consequence only if the relative increase of the groundwater flow is less than the relative increase of the rain- or meltwater flow when the discharge areas are expanding.

Below, the variation of the groundwater fraction with changes in the extent of the discharge area is discussed on the basis of three simple models for flow from hillslopes to the stream. The aim of the discussion is to find a simple conceptual model which is in keeping with the observed variations of the groundwater fraction: an inverse relationship between this fraction and total flow and a gradually decreasing slope in the diagrams showing this relationship, indicating less or no dependence of the groundwater fraction on total flow at moderate to high

discharge (Fig.8.5 within events and Fig.10.1 for the whole group of events). In this and the following discussions a basin is represented by a hillslope (or rather two opposite hillslopes) with the stream at the base. Flows are expressed per unit of width of the slope.

Let us assume that the soil in the three hillslopes sketched in Fig.10.4 is homogeneous and isotropic with respect to the saturated hydraulic conductivity. (The thickness of the till-soil cover generally decreases upslope. However, for the considerations below, such a variation exerts only little influence on the flow relationships obtained from the simplified slope a) in the figure.)

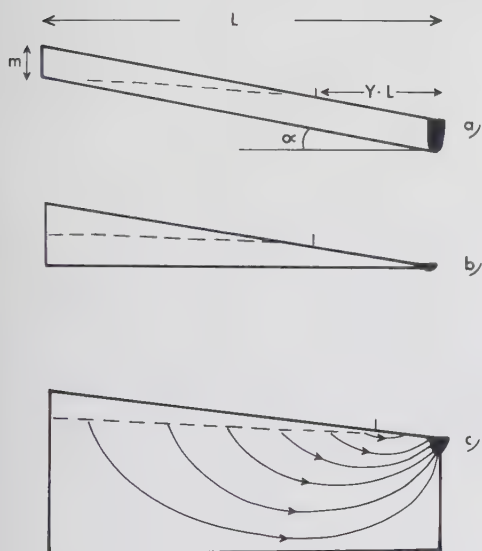


Fig.10.4 Hillslope models for discussion on the time variation of the groundwater fraction of streamwater.

Neglecting effects of storage, the rain- or meltwater flow to the stream, Q_p , is given by

$$Q_p = i Y L \quad (10.1)$$

where i is rainfall¹ intensity, Y the fraction of discharge area, and L the slope length.

For stationary flow conditions in both the unsaturated and saturated zones we get, regardless of slope geometry, the groundwater flow, Q_g , by

¹ In these discussions the term rainfall is used for the total water input to the ground, i.e., rainfall and snowmelt.

$$Q_g = i L (1 - Y) \quad (10.2)$$

The groundwater fraction, X , is then

$$X = \frac{Q_g}{Q_g + Q_p} = 1 - Y \quad (10.3)$$

i.e., X decreases linearly with increasing Y . The rainfall intensity, and thus the total flow ($i L$), is not seen directly in the relationship. The value of Y is, however, highly dependent on i (and on slope geometry and hydraulic properties). The larger the rainfall intensity, the larger Y will be, since the discharge area expands upslope to the point where the rainwater supply above, $i L (1-Y)$, equals the maximum possible groundwater flow through the section (see discussion in Eriksson, 1977, O'Loughlin, 1981, and Grip and Rodhe, 1985).

The above stationary conditions refer to long-term mean flow. The situation may be approached in a long snowmelt event, but is of little relevance for most of the events investigated here, mainly due to soil water storage. In the following discussion, stationary conditions are assumed only for the rainwater component (Eq.10.1). Applying the Dupuit assumption for groundwater flow in slopes a) and b) in Fig.10.4, the groundwater fraction can be related to rainfall intensity and the extent of discharge area. The maximum groundwater flow through a saturated section is then the transmissivity multiplied with the gradient of the ground surface.

For slope a) we have

$$Q_g = m K \tan \alpha \quad (10.4)$$

giving

$$X = \frac{1}{1 + \frac{i}{K m \tan \alpha} \cdot Y} = \frac{1}{1 + c_1 i Y} \quad (10.5)$$

Here m is the thickness of the conducting layer, K the saturated conductivity, α the slope angle and c_1 a constant.

In slope b) the maximum thickness of the saturated zone is $L Y \sin \alpha$, giving

$$Q_g = L Y K \sin \alpha \tan \alpha \quad (10.6)$$

and

$$X = \frac{1}{1 + \frac{i}{K \sin \alpha \tan \alpha}} = \frac{1}{1 + c_2 i} \quad (10.7)$$

where c_2 is a constant.

In slope c) the Dupuit assumption is no longer valid and the vertical components of flow have to be considered. With a horizontal groundwater table upslope from the saturated area, the groundwater flow is univocally determined by Y (given the geometrical configuration and K), showing a decreasing rate of increase with increasing Y . From flow calculations using the method of flow net, the dependence of Q_g on Y was found to be described by

$$Q_g \approx c_3 Y^n \quad (10.8)$$

where the exponent n is less than one. The value of c_3 represents the maximum groundwater flow through the slope, occurring when the slope is completely saturated. With increasing gradient of the water table above the discharge area, Q_g is larger at a certain value of Y and it approaches the maximum value more rapidly as Y increases. Using the approximate relationship of Eq.(10.8) we get

$$X = \frac{1}{1 + \frac{L}{c_3} i Y^{(1-n)}} \quad (10.9)$$

Of the three straight, homogeneous hillslopes, slope a) and slope b) constitute the two extremes of the variation of the groundwater flow with the extent of discharge area. In slope a) the groundwater flow reaches its maximum value as soon as the ground surface becomes saturated at the base. The effect of further increase of the groundwater level is just to expand the saturated area, thereby increasing the rainwater flow to the stream. Obviously, such a constant groundwater flow is not in accordance with observations. In slope b), on the other hand, the groundwater flow increases linearly with the saturated area. Since this is also the case for the rainwater flow, the

groundwater fraction is independent of Y . Slope c) falls somewhere in between slope a) and b). For small values of Y , the increase of Q_g with Y could as well be assumed to be linear, making X independent of Y also in this case.

Two severe objections can be raised against the application of these simplistic models to the present study basins. The hydraulic conductivity of the soil is not constant, but generally decreases with depth. Further, the slopes are not straight.

With slope a) a laterally layered soil would give the same expression for X (Eq.10.5), only that K then represents the mean conductivity. In slope b) decreasing conductivity with depth would give a gradually slower increase of Q_g with increasing Y . If the conductivity decreases exponentially with depth, Eq.(10.7) changes to

$$X = \frac{1}{1 + f(Y) \cdot i} \quad (10.10)$$

where $f(Y)$ is an increasing function of Y . If K decreases rapidly with the depth, $f(Y)$ is approaching $c_5 Y$ as Y approaches 1. The expression for X thus approaches case a) (considerable groundwater flow only in the top layer). The effect of the same conductivity variation would be similar in slope c).

With concave slopes, Q_g would decrease more rapidly with Y in all cases. The groundwater fraction would then show less rapid decrease with Y than in the straight slopes. In slope b) the groundwater fraction would even increase as Y increases, since Q_g would increase more rapidly than Q_p . In convex hillslopes the opposite would occur.

From the above discussion it can be concluded that the groundwater fraction is inversely related to the rainfall or snowmelt intensity by Eq.(10.10) with $f(Y)$ depending on the geometry and the hydraulic conductivity and its variation. In general $f(Y)$ is probably a constant or an increasing function of Y .

Eq.(10.10) does not account for the time lag between rainwater input and streamwater output. The equation implies, for instance, that X returns to 1.0 immediately after the cessation of the rainfall, which is not true. Therefore the equation is of little value when interpreting single rainfall events, in which the water input takes place during a short time as compared with the stream stormflow. But for the snowmelt events,

with long periods of water input, it may be helpful in interpreting the hydrograph separations.

With $f(Y)$ assumed to be a constant and i assumed to be proportional to total flow, Eq.(10.10) agrees qualitatively with the relationship found between groundwater fraction and total flow within the snowmelt events (Fig.8.4). The change of slope in the diagrams is accounted for by the appearance of the intensity, i , in the denominator. The hysteresis effect in the diagram (cf. Section 8.1.4) may be explained by the time lag between melting and streamflow. For the whole group of snowmelt events, Eq.(10.10) can be applied to the relationship found between volume fraction of groundwater and mean rate of water input (Fig.10.2). From Eq.(10.7) we have

$$f(Y) = c_2 = \frac{1}{K \sin \alpha \tan \alpha} \quad (10.11)$$

With the slope angle $\alpha = 10$ degrees (cf. Table 4.1a), the line drawn in Fig.10.2 corresponds to a saturated hydraulic conductivity of $2.4 \cdot 10^{-6}$ m/s. The value is somewhat low, but not unreasonable.

Let us stay with slope b) for a while and look at the extent of the discharge areas. Eq.(10.10), with $f(Y)$ according to Eq.(10.7), presumes a linear relationship between groundwater flow and the extent of discharge area. Above it was shown that the data for the melting periods could be roughly fitted to this model. Going in the opposite direction, it follows from the finding that the model can be applied to field data that the period means of discharge areas, as determined by ^{18}O , should be linearly related to the groundwater flow. Eq.(10.6) gives

$$Y = c_2 \cdot \frac{Q_g}{L} \quad (10.12)$$

with c_2 given by Eq.(10.11). Thus a relationship between Y and Q_g follows from the relationship found between groundwater fraction and rate of water input. The straight line in Fig.10.3 corresponds to $K = 2.4 \cdot 10^{-6}$ m/s for $\alpha = 10$ degrees.

10.2.4 Mechanisms for groundwater discharge response to water input

In the preceding section it was shown that the observed variations of groundwater fractions within events, and also between different events, may be in accordance with the present hypothesis for streamflow generation. The discussion was based on the variations of the discharge areas and the corresponding variations of groundwater flow. No attempts were made to explain the dynamics of the flow situations. In this section, the mechanisms responsible for the observed rapid increase in groundwater flow to the stream following a water input will be discussed.

The view formulated by Horton (1933), with overland flow due to rainfall excess as the dominating component of stormflow in streams, has had a very powerful impact on the conception of streamflow generation in hydrology and water engineering. When chemical and isotope studies, in the late 1960s, showed that stormflow was often dominated by pre-event water, researchers were perplexed. How can groundwater flow, known as a "slow" process, react so quickly to rain or meltwater input? The question has not yet received a comprehensive answer and the statement of Martinec (1975), restated by Sklash and Farvolden, (1979),

The exact nature of this propagation is not known,

is still valid. Various processes have been suggested: lateral unsaturated flow with a saturated zone at the base of the slopes acting as a "conduit" (Hewlett and Hibbert, 1967), groundwater wave propagation (Martinec, 1975), macropore flow (Mosley, 1979) and rapid head increase due to the effect of the capillary fringe (Sklash and Farvolden, 1979; Gillham, 1984). Below, it will be argued that in till soils, the existence of a large increase in hydraulic conductivity towards the ground surface may give one explanation for the observed rapid increases in groundwater discharge. In a homogeneous soil, rapid head increments by the effect of the capillary fringe may be important.

For the observed large and rapid response of the groundwater discharge to water input, two conditions must be fulfilled in the discharge area:

1. The hydraulic conductivity and the cross section for flow must be large enough.

2. A sufficiently large gradient in total head must build up rapidly enough.

These factors will be examined below on the basis of the hillslopes in Fig.10.4.

Let us first assume that the extent of the discharge area is constant over time. Then the relative increase in groundwater discharge is proportional to the relative change in head gradient over the discharging surface. In a typical runoff event, the groundwater peakflow is some 5 to 50 times the pre-event value, say 10 times. A corresponding relative increase in head gradient is needed to explain the flow variations. As mentioned earlier, however, the water table in till soils largely follows the ground surface, with the depth increasing from a few decimetres at the base to a few metres at the top of the hillslopes. In a straight slope, having the streamwater level at the ground surface or a saturated discharge area at the base, the maximum slope of the water table is that of the ground surface. Under such conditions, the relative increase in the slope of the water table in a typical hillslope will not exceed some 10%, giving the same relative increase in groundwater flow. Head increments alone are thus far from capable of generating the observed many-fold increase in groundwater discharge.

Let us now allow the extent of the discharge area to vary. In Section 10.2.3 it was shown that the relative increase in groundwater discharge could be about the same as the relative increase in discharge area (slope b), homogeneous), or even exceed it (slope b), concave). A some 10-fold relative increase in the extent of the discharge area within events seems high, but is not fully unrealistic. Situations can thus be imagined where an increase in the hydraulic capacity of the discharge area could explain the increase in groundwater flow, even when the hydraulic conductivity is constant with depth within the conducting layer.

A more realistic case is when the hydraulic conductivity varies with depth. From the observed depth variation of the hydraulic conductivity Lundin (1982) established a groundwater rating curve for constant slope of the water table, here presented in a slightly different way in Fig.10.5. According to this figure, a rise in the groundwater level from 0.6 to 0.1 m depth could readily explain the variations in groundwater flow.

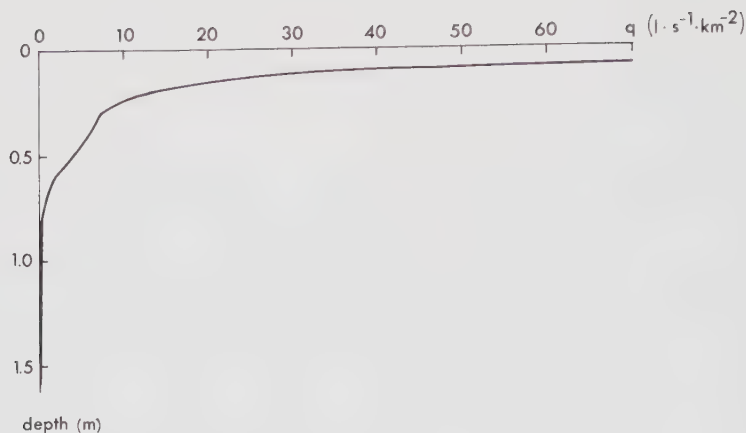


Fig.10.5 Total groundwater flow as a function of the depth to the groundwater table calculated from the hydraulic conductivity profile. The hillslope is assumed to be 50 m long and the slope of the water table 0.1 m/m at the base. Masbybäcken (after Lundin, 1982).

We have seen that, with the above models, the relative changes in head gradient are too small to explain observed changes in groundwater discharge from hillslopes with homogeneous soil. With constant hydraulic conductivity, a large increase in the cross-sectional area for flow, i.e., in the extent of the discharge area, could explain the changes, but this is less probable. With the conductivity increasing towards the ground surface, the variation in the groundwater discharge may, on the other hand, well be related to variations in the groundwater level. It can also be noted that coarse layers which may occur at depth in till soil (e.g. Knutsson, 1971) or in the border between the till soil and the bedrock, contribute little to the increase in groundwater discharge. This is so, since the relative changes in head gradient along such a lense are small as long as the water table is continuously above the lense.

Let us now add a time aspect and discuss the conditions for a sufficiently rapid and large head increment to build up.

The rise of the groundwater table that follows upon a certain infiltration is closely related to the unsaturated conductivity and the water-holding capacity of the soil in the unsaturated zone (which in turn are closely related). The conductivity determines the rate of pressure propagation, i.e. the time of

response, whereas the water-holding capacity governs the magnitude of the rise. In a homogeneous soil the water table will show the quickest response close to the discharge area, where the largest rise of the water table often takes place as well. The high pressure propagation velocity in the unsaturated zone, due to comparatively high water content under the moderate tensions prevailing, and the short distance to the water table, gives a rapid response. The high water content in the whole unsaturated zone further causes a small amount of added water to produce a large rise of the water table, i.e., the storage coefficient is small. Due to this rapid and large response in areas with very shallow groundwater, groundwater ridges may develop close to the stream, as observed in the field by Ragan (1968) and in mathematical modelling of hillslope flow (e.g., Johansson, 1985).

The above effects are the basis of the capillary fringe concept, advocated as an explanation for the rapid response of groundwater discharge observed from ^{18}O by Sklash and Farvolden (1979). In areas where the depth to the water table is less than the air entry pressure of the water retention curve (the so-called capillary fringe), small volumes of infiltrated water may generate very rapid and large head increments, shown theoretically with some field support by Gillham (1984). With suitable geometry of the stream channel side, the head gradient towards the stream, and thus the groundwater outflow, may increase many-fold, as shown in laboratory experiments by Abdul and Gillham (1984). The capillary fringe concept may thus offer an explanation for the observed increase of groundwater outflow also in soils with little or no increase in the hydraulic conductivity towards the ground surface.

The capillary fringe concept may also explain the occurrence of "deep groundwater stormflow", the existence of which was indicated from hydrochemical observations by Calles (1982) (cf. Section 2.3). Large raisings of the water table close to the stream or close to a low-lying area can be generated due to the effects of the capillary fringe. Such raisings are a necessary condition for a substantial increase in the vertical outflow of deep groundwater. Large rises in the water table higher up in the hillslope can give but small relative changes in the vertical head gradient towards the stream or its surrounding discharge area.

One necessary condition for the capillary fringe concept to be applicable in streamflow generation is that the soil has little

or no pore volume air filled at tensions smaller than, say, 0.3 m. For sorted soils, without structure, this is fulfilled for fine sand and more fine-grained soils (Andersson and Wiklert, 1972). The effect of pore size on streamflow generation contains two contradictory effects: In a coarse soil the saturated hydraulic conductivity is great and substantial flow can take place, but the water table is often deep and the capillary fringe small, counteracting a rapid response of groundwater discharge upon infiltration. In a fine soil, on the other hand, the head may react rapidly and profoundly, but the possible flow rates are restricted due to the low hydraulic conductivity.

In a till soil, having pores of very different sizes, the pores empty more gradually by increased tension (= more negative pressure potential). The air entry pressure is often a small negative number (a few centimetres), particularly in the shallow layers where there may be large structural pores due to various soil processes. This is seen from the retention curves by Lundin (1982) and Johansson (1986), which show a systematic increase in the air entry pressure (= getting less negative) as the ground surface is approached. Such a variation counteracts rapid and large head increments. But since the hydraulic conductivity increases with height, a large increment in the head gradient is not necessary for a large increase in the groundwater discharge to take place. The mathematical model experiments in a layered hillslope by Johansson (1985) (cf. Section 2.2) showed that moderate increments in the head gradient and in the cross section for flow close to the stream were sufficient to generate groundwater flow hydrographs similar to observed streamflow hydrographs.

10.2.5 The origin of stormflow

This thesis has experimentally treated one aspect of the runoff process: the origin of streamflow. The word "origin" has been used very generally to denote which type of water particles contribute to stream stormflow, new rain- or meltwater on the one hand or groundwater on the other.

In calculating the extent of discharge areas by Eq.(3.9), all new rain- or meltwater on the discharge area was assumed to be discharged by the stream during the event. This assumption agrees with the velocity of overland flow to be expected, a few centimetres per second, implying that overland flowing-water

from the whole discharge area reaches the stream within minutes or hours.

No measurements have been made in the study to investigate from where in the basin the discharging groundwater particles originate. The particle velocity for flow in the soil is smaller than that for overland flow by several orders of magnitude. The vertical velocity in the unsaturated zone seldom exceeds a few cm/h, and this velocity remains only during and shortly after the time of infiltration (cf. the mean particle velocities around 1 m/year observed by Saxena, 1987). It is thus only in areas with very shallow groundwater that rainwater particles reach the groundwater during the runoff event. Rough estimates, based on observed hydraulic conductivities and porosities in till soils, show that the groundwater particles discharged during an event, were probably located within or in close vicinity of the discharge area at the onset of the event.

The above comments concern the origin of the water particles. Below, the origin of stormflow will be discussed from a different angle: Where in the basin are the impulses for stormflow given? In these discussions, a sub-area is said to contribute if rainfall or snowmelt in the sub-area during a runoff event in some way affects the streamflow during the event, i.e., if the water input gives impulses for stream stormflow.

Generally speaking, the impulses from the water input to stormflow grow weaker the longer the distance to the stream is.

Saturated discharge areas (including the stream) can be said to contribute 100%. Ignoring the evapotranspiration during the flow event, the flow impulse from these areas equals the rainfall or snowmelt they receive. Here the impulses are propagated by saturation overland flow.

Due to storage and evapotranspiration, the volume discharged to the stream as a result of infiltration in the recharge areas is smaller than the total infiltrated volume. Thus recharge areas do not contribute 100%. As pointed out above, the lowest parts of the recharge areas, where the groundwater table is very shallow and the soil water content high, play a key role in the streamflow generation. In these edge areas of the discharge area, the contribution to streamflow can be close to 100%, but upslope it gradually decreases with increasing distance to the discharge area. In areas where the water flow is concentrated

by the topography, i.e., in topographic hollows and other concave landforms, the impulses to stormflow are particularly large, by saturation overland flow as well as by rapid groundwater discharge response to infiltration.

The basin fraction of (saturated) discharge areas, Y , was estimated by Eq.(3.9), which can be written

$$Y = V_p/V_{wi} \quad (10.13)$$

where V_p is the discharged volume of new rain- or meltwater and V_{wi} the volume of the water input to the basin, with both volumes expressed for instance as mm of water depth over the basin. The extent of areas giving impulses for groundwater discharge cannot be estimated in a corresponding way, however, since these areas do not contribute 100%. But we know that these areas are larger than the area needed if the contribution were 100%, i.e., an area equal to V_g/V_{wi} , where V_g is the discharged volume of groundwater. Since the runoff events are generally dominated by groundwater ($V_g > V_p$), the impulses must come from an area considerably larger than the (saturated) discharge area. The minimum area needed to generate a runoff event is $(V_p + V_g)/V_{wi}$, the area needed with a 100% contribution (the so-called minimum contributing area discussed by Dickinson and Whiteley, 1970). Expressed as a fraction of the basin area, this area equals the discharged fraction of the water input. In this study, this fraction varied from 10% in small summer events to more than 90% in large autumn or spring floods (Tables A2.1 and A3.1).

10.2.6 Transit times for water in a hillslope

The estimated reservoir volumes for water in the basins, commented upon in Section 10.1.3, correspond to mean transit times for water of around 0.5-1.5 year. In this final section, the transit time distribution for water discharging from a hypothetical hillslope is estimated from flow considerations. The estimates are based on the time frequency distribution for groundwater levels observed by Lundin (1982). These distribution curves refer to spring and summer periods, but are here assumed to represent the whole year.

The flow in the saturated zone is assumed to be lateral, driven by a slope of the water table of 0.1. The long-term mean flow rate at the depth z , $q(z)$, is calculated as

$$q(z) = K(z) \frac{dh}{dx} f(z) \quad (10.14)$$

where dh/dx is the slope of the groundwater table and $f(z)$ the fraction of time with the groundwater level at or above the level. Fig.10.6 shows the accumulated mean discharge (per unit of width) below the depth z , $Q(z)$, calculated as

$$Q(z) = \int_z^{z_0} q(z) dz \quad (10.15)$$

where z_0 is the depth of the flow region¹.

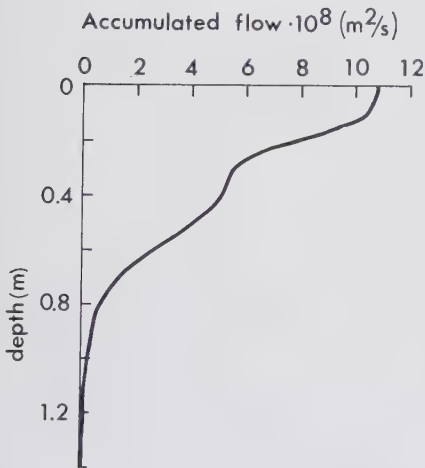


Fig.10.6 Accumulated long-term mean groundwater flow below different depths in the soil. Masbybäcken (data from Lundin, 1982).

The mean groundwater recharge, R , is assumed to be constant along the slope. It is further assumed that the deepest groundwater is recharged at the top of the hillslope and that the shallowest groundwater at the base is recharged at the base. For groundwater flowing at the level z_x at the base of the slope, we then get the distance from the base to the point of recharge, expressed as a fraction, x , of the hillslope length, L , from

$$R (1 - x) L = \int_{z_x}^{z_0} q(z) dz \quad (10.16)$$

¹ For reasonable values of slope lengths, the yearly runoff according to Fig.10.6 is smaller than those observed. This discrepancy is disregarded here, however, since it is the depth variation of flow that is of interest.

Since the total flow at the base of the slope is

$$R L = \int_{z_{wt}}^{z_o} q(z) dz \quad (10.17)$$

we get

$$1 - x = \frac{\int_{z_x}^{z_o} q(z) dz}{\int_{z_{wt}}^{z_o} q(z) dz} \quad (10.18)$$

where z_{wt} is the depth of the highest observed water table. With this equation, z_x is obtained graphically from Fig.10.6. The transit time for the groundwater recharged at level x is then obtained as

$$t_x = \frac{L x}{v_p(z_x)} = \frac{L x p(z_x)}{K(z_x) \frac{dh}{dx}} \quad (10.19)$$

where v_p is the particle velocity of groundwater and p is the porosity. The transit time is plotted in Fig.10.7 (broken line). Due to the rapid decline in conductivity with depth, the effect of which is added to the longer distance for water recharged upslope, the distribution curve for groundwater is spread out. A considerable fraction of the groundwater thus has short transit times, 20% < 10 days, but there is also a considerable fraction of groundwater having long transit times, 10% > 2 years.

The transit times in the unsaturated zone are estimated assuming vertical flow with a particle velocity of 1 m/year (cf. Saxena, 1987). The thickness of the unsaturated zone is assumed to vary linearly from 0 m at the base to 1.5 m at the top of the hill-slope.

In Fig.10.7 the total transit times for water in the hillslope, estimated as the sum along the flow paths of those calculated for unsaturated and saturated flow, are shown (unbroken line). With an assumed slope length of 50 m, the median total transit time is about 300 days, with the median transit time for groundwater being about 100 days. The very long transit times have a

large influence upon the mean transit time, which in this slope is some 1-2 years if no flow is assumed below 1.5 m depth. Regarding the actual two-dimensional flow pattern of the groundwater, causing still smaller flow rates and longer flowpaths of the deep groundwater, the true mean transit time of groundwater discharging to the stream may be substantially longer.

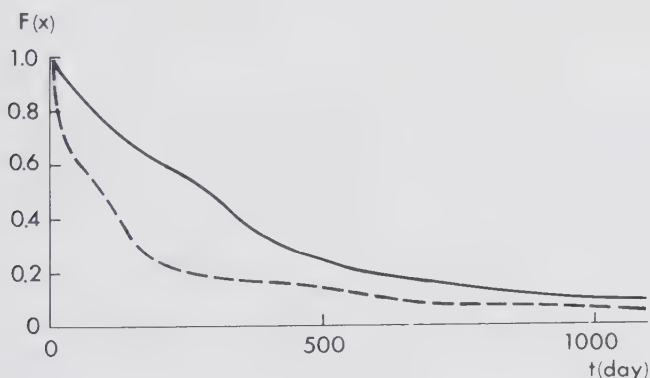


Fig.10.7 Transit time distribution for groundwater only (broken line) and total transit times (unbroken line) for flow in a 50 m hillslope, estimated from conductivity profile and frequency distribution of groundwater levels. Masbybäcken (data from Lundin, 1982).

When making similar estimates for stationary conditions representing mean flow rate in the hillslope, the median transit time for groundwater becomes longer, since the rapid flow of short duration in the high-conductivity surface layer is not then included. The long tail of the distribution curve, on the other hand, remains similar.

It should be noted that the estimates in this section are based on conductivity profiles in one single basin, namely the neighbour basin of one of the present study basins, and that the geometric conditions and the flow patterns are highly simplified. One important question concerns the relationship between flow rate and particle velocity for groundwater. It has been assumed here that all water in the soil contributes to the flow, (Eq.10.19). If there is a large fraction of "immobile" water in the pores, the particle velocities are larger and the transit times shorter. Another important factor is a possible occurrence of high conductivity layers at depth, either in the soil or in the underlying bedrock or in the border between them.

Such layers, which are probably common in many till soils, would act as drains for the layers above, and could significantly decrease the transit times during saturated flow. Nevertheless, the transit times estimated here from simple flow observations agree qualitatively with those obtained by ^{18}O in this and other studies.

REFERENCES

- Aastrup, M., Aneblom, T., Henriksson, B. and Persson, G. (1982): PMK-grundvatten. SGU, Rapporter och meddelanden 28, 45 pp.
- Abdul, A.S. and Gillham, R.W. (1984): Laboratory studies of the effects of the capillary fringe on streamflow generation. Water Resour. Res., 20(6):691-698.
- Allison, G.B. and West, S.F. (1979): Estimation of catchment infiltration capacity. Proc. Symp. Hydrology and Water Resources, Perth 10-12 Sept., Inst. of Engineers, Perth, 64-68.
- Allison, G.B. (1982): The relationship between ^{18}O and deuterium in water in sand columns undergoing evaporation. J. Hydrol., 55:163-169.
- Allison, G.B. and Leaney, F.W. (1982): Estimation of isotopic exchange parameters, using constant-feed pans. J. Hydrol., 55:151-161.
- Allison, G.B., Barnes, C.J., Hughes, M.W. and Leaney, F.W.J. (1984): Effect of climate and vegetation on oxygen-18 and deuterium profiles in soils. In: Isotope hydrology 1983. IAEA, Vienna 1984:105-123.
- Ambach, W., Elsässer, M., Moser, H., Rauert, W., Stichler, W. and Trimborn, P. (1975): Variationen des Gehalts an Deuterium, Sauerstoff-18 und Tritium während einzelner Niederschläge. Wetter und Leben, 27:186-192.
- Anderson, M.G. and Burt, T.P. (1978a): Experimental investigations concerning the topographic control of soil water movement on hillslopes. Z. Geomorph. N.F., Suppl. Bd. 29:52-63.
- Anderson, M.G. and Burt, T.P. (1978b): Toward more detailed field monitoring of variable source areas. Water Resour. Res., 14(6):1123-1131.
- Anderson, M.G. and Kneale, P.E. (1980): Topography and hillslope soil water relationships in a catchment of low relief. J. Hydrol., 47:115-128.
- Anderson, M.G. and Burt, T.P. (1982): The contribution of throughflow to storm runoff: an evaluation of a chemical mixing model. Earth Surface Processes and Landforms, 7:565-574.
- Anderson, M.G. and Burt, T.P. (eds) (1985): Hydrological forecasting. John Wiley & Sons, Chichester, 604 pp.
- Andersson, S. and Wiklert, P. (1972): Markfysikaliska undersökningar i odlad jord. XXIII. Om de vattenhållande egenskaperna hos svenska jordarter. Grundförbättring (Journal of Agricultural Land Improvement), 25(2-3):53-143.
- Ando, Y., Musiak, K. and Takahasi, Y. (1983): Modelling of hydrologic processes in a small natural hillslope basin, based on the synthesis of partial hydrological relationships. J. Hydrol., 64:311-337.
- Appelo, C.A.J., Becht, R., Griend, A.A. Van de, and Spierings, T.C.M. (1983): Build-up of discharge along the course of a mountain stream deduced from water-quality routings (EC routings). J. Hydrol., 66:305-318.

- Arnason, B. (1969): The exchange of hydrogen isotopes between ice and water in temperate glaciers. Earth Planet. Sci. Lett., 6:423-430.
- Arnason, B., Buason, Th., Martinec, J. and Theodorson, P. (1973): Movement of water through snowpack traced by deuterium and tritium. In: The role of snow and ice in hydrology. Proc. Banff Symp., Sept. 1972, UNESCO-WMO-IAHS, IAHS Publ. 107:299-312.
- Arnason, B. (1981): Ice and snow hydrology. In: Stable isotope hydrology. Deuterium and oxygen-18 in the hydrologic cycle. IAEA, Technical reports series No 210, Vienna, 143-175.
- Balek, J., Moldan, B., Paces, T. and Skorepa, J. (1978): Hydrological and geochemical mass balance in small forested and agricultural basins. Proc. Symp. Baden, Sept. 1978, IAHS Publ. 125:50-58.
- Beasley, R.S. (1976): Contribution of subsurface flow from the upper slopes of forested watersheds to channel flow. Soil Sci. Soc. Am. J., 40:955-957.
- Behrens, H., Bergmann, H., Moser, H., Rauert, W., Stichler, W., Ambach, W., Eisner, H. and Pessl, K. (1971): Study of the discharge of alpine glaciers by means of environmental isotopes and dye tracers. Z. Gletscherkd. Glazialgeol., 7(1-2):79-102.
- Behrens, H., Moser, H., Oerter, H., Rauert, W. and Stichler, W. (1979): Models for the runoff from a glaciated catchment area using measurements of environmental isotope contents. In: Isotope hydrology 1978. IAEA, Vienna, 829-846.
- Bergqvist, E. (1971a): Nästen and Marsta - two small drainage basins in central Sweden, 1. Purpose of the investigation and description of the basins. UNGI Report 5, Uppsala Univ., Dept Phys. Geogr., 49 pp.
- Bergqvist, E. (1971b): Nästen and Marsta - two small drainage basins in central Sweden, 4. Runoff and discharge of suspended matter and dissolved solids at Stabby and Sundbro. UNGI Report 8, Uppsala Univ., Dept Phys. Geogr., 123 pp.
- Bergström, S. and Sandberg, G. (1983): Simulation of groundwater response by conceptual models - Three case studies. Nord. Hydrol., 14:71-84.
- Betson, R.P. (1964): What is watershed runoff? J. Geophys. Res., 69(8): 1541-1551.
- Betson, R.P., Marius, J.B. and Joyce, R.T. (1968): Detection of saturated interflow in soils with piezometers. Soil Sci. Soc. Am. Proc., 32: 602-604.
- Beven, K. (1977): Hillslope hydrographs by the finite element method. Earth Surface Processes, 2:13-28.
- Beven, K. (1978): The hydrological response of headwater and sideslope areas. Hydrol. Sci. Bull., 23(4):419-437.
- Beven, K. (1981): Kinematic subsurface stormflow. Water Resour. Res., 17(5):1419-1424.
- Beven, K. (1982a): On subsurface stormflow: an analysis of response times. Hydrol. Sci. J., 4(12):505-521.

- Beven, K. (1982b): On subsurface stormflow: Predictions with simple kinematic theory for saturated and unsaturated flows. Water Resour. Res., 18(6):1627-1633.
- Blavoux, B. and Siwertz, E. (1971): Traçage isotopique de l'infiltration sur lysimètre. C.R. Acad. Sci. Paris, 273:2056-2059.
- Blavoux, B. (1978): Étude de cycle de l'eau au moyen de l'oxygène 18 et du tritium. Possibilités et limites de la méthode des isotopes du milieu en hydrologie de la zone tempérée. Thèse de doctorat d'État, Université Pierre et Marie Curie, Paris.
- Bleeker, W., Dansgaard, W. and Lablans, W.N. (1966): Some remarks on simultaneous measurements of particulate contaminants including radioactivity and isotopic composition of precipitation. Tellus, 18: 773-785.
- Bonell, M. and Gilmour, D.A. (1978): The development of overland flow in a tropical rainforest catchment. J. Hydrol., 39:365-382.
- Bonell, M., Hendriks, M.R., Imeson, A.C. and Hazelhoff, L. (1984): The generation of storm runoff in a forested clayey drainage basin in Luxembourg. J. Hydrol., 71:53-77.
- Bottomley, D.J., Craig, D. and Johnston, L.M. (1984): Neutralization of acid runoff by groundwater discharge to stream in Canadian precambrian shield watersheds. J. Hydrol., 75:1-26.
- Bottomley, D.J., Craig, D. and Johnston, L.M. (1986): Oxygen-18 studies of snowmelt runoff in a small precambrian shield watershed: Implications for streamwater acidification in acid-sensitive terrain. J. Hydrol., 88: 213-234.
- Bren, L.J. (1980): Hydrology of a small, forested catchment. Australian For. Res., 10:39-51.
- Bren, L.J. and Turner, A.K. (1985): Hydrologic behaviour of a small forested catchment. J. Hydrol., 76:333-350.
- Bringfeldt, B. and Hårsmar, P-O. (1974): Rainfall interception in a forest in the Velen hydrological representative basin. Nord. Hydrol., 5(3): 146-165.
- Buason, Th. (1972): Equations of isotope fractionation between ice and water in a melting snow column with continuous rain and percolation. J. Glaciol., 11:387-405.
- Burgman, J.O., Calles, B. and Westman, F. (1987): Conclusions from a ten year study of oxygen-18 in precipitation and run-off in Sweden. Proc. Symp. Use of Isotope Techniques in Water Resources Development, IAEA, Vienna, 30 March-3 April 1987 (in press).
- Calles, U.M. (1982): Diurnal variations in electrical conductivity of water in a small stream. Nord. Hydrol., 13:157-164.
- Calles, U.M. (1985): Deep groundwater contribution to a small stream. Nord. Hydrol., 16(1):45-54.
- Carslaw, H.S. and Jaeger, J.C. (1959): Conduction of heat in solids. 2nd ed., Oxford Univ. Press, London, 510 pp.

- Christophersen, N., Kjaernsrod, S. and Rodhe, A. (1984): Preliminary evaluation of flow patterns in the Birkenes catchment using natural ^{18}O as a tracer. In: Johansson, I. (ed.), Hydrological and hydrogeochemical mechanisms and model approaches to the acidification of ecological systems. Rep. 10, Nord. Hydrol. Progr., Oslo, 29-40.
- Corbett, E.S., Sopper, W. and Lynch, J.A. (1975): Watershed response to partial area applications of simulated rainfall. IAHS Publ. 117:63-73.
- Craig, H. (1961a): Standard for reporting concentrations of deuterium and oxygen-18 in natural waters. Science, 133:1833-1834.
- Craig, H. (1961b): Isotopic variations in meteoric waters. Science, 133:1702-1703.
- Craig, H. and Gordon, L. (1965): Deuterium and oxygen-18 variation in the ocean and the marine atmosphere. In: Tongiorgi, E. (ed.), Stable isotopes in oceanographic studies and paleo-temperatures. Spoleto, 1965, 9-130.
- Crouzet, E., Hubert, P., Olive, Ph. and Siwertz, E. (1970): Le tritium dans les mesures d'hydrologie de surface. Détermination expérimentale du coefficient de ruissellement. J. Hydrol., 11:217-229.
- Dansgaard, W. (1961): The isotopic composition of natural waters with special reference to the Greenland ice cap. Meddelelser om Grönland, 165(2):1-120.
- Dansgaard, W. (1964): Stable isotopes in precipitation. Tellus, 16:436-468.
- Dickinson, W.T. and Whiteley, H. (1970): Watershed areas contributing to runoff. IAHS Publ. 96:1.12-1.26.
- Dinçer, T., Payne, B.R., Florkowski, T., Martinec, J. and Tongiorgi, E. (1970): Snowmelt runoff from measurements of tritium and oxygen-18. Water Resour. Res., 6:110-124.
- Dinçer, T., Al-Mugrin, A. and Zimmermann, U. (1974): Study of the infiltration and recharge through the sand dunes in arid zones with special reference to the stable isotopes and thermonuclear tritium. J. Hydrol., 23:79-109.
- Dunne, T. and Black, R.D. (1970a): An experimental investigation of runoff production in permeable soils. Water Resour. Res., 6(2):478-490.
- Dunne, T. and Black, R.D. (1970b): Partial area contributions to storm runoff in a small New England watershed. Water Resour. Res., 6(5):1296-1311.
- Dunne, T. and Black, R.D. (1971): Runoff processes during snowmelt. Water Resour. Res., 7(5):1160-1172.
- Dunne, T., Moore, T.R. and Taylor, C.H. (1975): Recognition and prediction of runoff-producing zones in humid regions. Hydrol. Sci. Bull., 20:305-327.
- Dunne, T. (1983): Relation of field studies and modeling in the prediction of storm runoff. J. Hydrol., 65:25-48.
- Eichler, R. (1965): Deuterium-Isotopengeochemie des Grund- und Oberflächenwassers. Geol. Rundsch., 55:144-155.

- Epstein, S. and Sharp, R.P. (1959): Oxygen isotope variations in the Malaspina and Saskatchewan glaciers. J. Geol., 67:88-102.
- Eriksson, B. (1983): Data rörande Sveriges nederbördsclimat, normalvärden för perioden 1951-80. SMHI, Rapport 28, 91 pp.
- Eriksson, E. (1961): The exchange of matter between atmosphere and sea. In: Oceanography. Publ. No 67 of the Am. Ass. Adv. Sc., 411-423.
- Eriksson, E. (1965): Deuterium and oxygen-18 in precipitation and other natural waters. Some theoretical considerations. Tellus, 17:498-512.
- Eriksson, E. (1977): Ground-water recharge and surface run-off. STRIAE, 4:11-12.
- Eriksson, E. (1985): Principles and applications of hydrochemistry. Chapman and Hall, London, 187 pp.
- Freeze, R.A. and Witherspoon, P.A. (1967): Theoretical analysis of regional groundwater flow, 2. Effect of watertable configuration and subsurface permeability variation. Water Resour. Res., 3:623-634.
- Freeze, R.A. (1971): Three-dimensional, transient, saturated-unsaturated flow in a groundwater basin. Water Resour. Res., 7(5):1160-1172.
- Freeze, R.A. (1972a): Role of subsurface flow in generating surface runoff, 1. Baseflow contributions to channel flow. Water Resour. Res., 8(3): 609-623.
- Freeze, R.A. (1972b): Role of subsurface flow in generating surface runoff, 2. Upstream source areas. Water Resour. Res., 8(5):1272-1283.
- Freeze, R.A. (1974): Streamflow generation. Rev. Geophys. Space Phys., 12(4):627-647.
- Freeze, R.A. and Cherry, J.A. (1979): Groundwater. Prentice-Hall, Inc., Englewood Cliffs, New Jersey, 604 pp.
- Freeze, R.A. (1980): A stochastic-conceptual analysis of rainfall-runoff processes on a hillslope. Water Resour. Res., 16:391-408.
- Fritz, P., Cherry, J.A., Weyer, K.V. and Sklash, M.G. (1976): Runoff analyses using environmental isotopes and major ions. In: Interpretation of environmental isotope and hydrochemical data in groundwater hydrology. IAEA, Vienna, 111-130.
- Gat, J.R. and Tzur, Y. (1967): Modification of the isotopic composition of rainwater by processes which occur before groundwater recharge. In: Isotopes in hydrology. IAEA, Vienna, 49-60.
- Gat, J.R. (1970): Environmental isotope balances of Lake Tiberias. In: Isotope hydrology. IAEA, Vienna, 109-120.
- Gillham, R.W. (1984): The capillary fringe and its effect on water-table response. J. Hydrol., 67:307-324.
- Glover, B.J. and Johnson, P. (1974): Variations in the natural concentration of river water during flood flows, and the lag effect. J. Hydrol., 22:303-316.

- Gonfiantini, R., Gratzin, S. and Tongiorni, E. (1965): Oxygen isotopic composition of water in leaves. In: Isotopes and radiation in soil-plant nutrition studies. IAEA, Vienna, 405-409.
- Gotthardsson, M. (1973): Jorddjupskartering i Stormyra. SMHI, Rapport 11.
- Granger, R.J., Gray, D.M. and Dyck, G.E. (1984): Snowmelt infiltration to frozen Prairie soils. Can. J. Earth Sci., 21:669-677.
- Griend, A.A., Van de, and Arwert, J.A. (1983): The mechanism of runoff generation from an alpine glacier during a storm traced by oxygen $^{18}\text{O}/^{16}\text{O}$. J. Hydrol., 62:263-278.
- Grip, H. (1982): Water chemistry and runoff in forest streams at Kloten. UNGI Report 58, Uppsala Univ., Dept Phys. Geogr., 144 pp.
- Grip, H. and Rodhe, A. (1985): Vattnets väg från regn till bäck. Forskningsrådets förlagstjänst, Stockholm, 156 pp.
- Gustafsson, Y. (1946): Die Strömungsverhältnisse in gedräntem Boden. Acta Agriculturae Suecana, 2(1):1-157.
- Gustafsson, Y. (1968): The influence of topography on groundwater formation. In: Eriksson, E., Gustafsson, Y. and Nilsson, K. (eds), Ground water problems. Pergamon Press, Oxford, 3-21.
- Handbook of chemistry and physics (56th ed.). Weast, R.C. (ed.), CRC Press, Cleveland, Ohio, 1975.
- Harr, R.D. (1977): Water flux in soil and subsoil on a steep forested slope. J. Hydrol., 33:37-58.
- Haupt, H.F. (1967): Infiltration, overland flow, and soil movement on frozen and snow-covered plots. Water Resour. Res., 3(1):145-161.
- Heede, B.H. (1984): Overland flow and sediment delivery: an experiment with small subdrainage in southwestern Ponderosa pine forests (Colorado, U.S.A.). J. Hydrol., 72:261-273.
- Helvey, J.D., Hewlett, J.D. and Douglass, J.E. (1972): Predicting soil moisture in the southern Appalachians. Soil Sci. Soc. Am. Proc., 36: 954-959.
- Henderson, F.M. (1966): Open channel flow. Macmillan, New York, 365-367, 450-472.
- Henninger, D.L., Petersen, G.W. and Engman, E.T. (1976): Surface soil moisture within a watershed - variations, factors influencing, and relationship to surface runoff. Soil Sci. Soc. Am. J., 40:773-776.
- Herrmann, A. (1978): A recording snow lysimeter. J. Glaciology, 82(20): 209-231.
- Herrmann, A., and Stichler, W. (1978): Variations d'isotopes stables dans une couche de neige alpine et leur application aux études hydrologiques. In: 2ème rencontre sur la neige et les avalanches de l'ANENA, Grenoble, 12 - 14 avril 1978, 81-91.
- Herrmann, A., Martinec, J. and Stichler, W. (1978): Study of snowmelt-runoff components using isotope measurements. Proc. Modeling of Snow Cover Runoff, U.S. Army Cold Regions Research and Engineering Laboratory, Hanover, New Hampshire, 26-28 Sept. 1978, 288-296.

- Herrmann, A. and Stichler, W. (1980): Groundwater - runoff relationships. Catena, 7:251-263.
- Herrmann, A., Lehrer, M. and Stichler, W. (1981): Isotope input into runoff systems from melting snow covers. Nord. Hydrol., 12:309-318.
- Hesmer, H. and Feldmann, A. (1953): Der Oberflächenabfluss auf bewaldeten und unbewaldeten Hangflächen des südlichen Sauerlandes. Forstarchiv, 24(11/12):245-256.
- Hewlett, J.D. (1961): Soil moisture as a source of base flow from steep mountain watersheds. US. Dept Agric. Forest Ser., Southeastern Forest Experiment Station, Asheville, North Carolina, Station Paper 132, 11 pp.
- Hewlett, J.D. and Hibbert, A.R. (1963): Moisture and energy conditions within a sloping soil mass during drainage. J. Geophys. Res., 68(4): 1081-1087.
- Hewlett, J.D. and Hibbert, A.R. (1967): Factors affecting the response of small watersheds to precipitation in humid areas. In: Sopper, W.E. and Lull, H.W. (eds), Int. Symp. on Forest Hydrology 1965, Penn. State Univ., Pergamon press, New York, 275-290.
- Hewlett, J.D. (1974): Comments on letters relating to Role of subsurface flow in generating surface runoff, 2. Upstream source areas by R. Allan Freeze. Water Resour. Res., 10(3):605-607.
- Hewlett, J.D., Fortson, J.C. and Cunningham, G.B. (1977): The effect of rainfall intensity on storm flow and peak discharge from forest land. Water Resour. Res., 13(2):259-266.
- Hills, R.C. (1971): The influence of land management and soil characteristics on infiltration and the occurrence of overland flow. J. Hydrol., 13:163-181.
- Hooper, R.P. and Shoemaker, C.A. (1986): A comparison of chemical and isotopic hydrograph separation. Water Resour. Res., 22(10):1444-1454.
- Hoover, M.D. and Hursh, C.R. (1943): Influence of topography and soil-depth on runoff from forest land. Trans. Am. Geophys. Union, 24:692-698.
- Horton, R.E. (1933): The rôle of infiltration in the hydrological cycle. Trans. Am. Geophys. Union, 14:446-460.
- Horton, R.E. (1945): Erosional development of streams and their drainage basins: hydrophysical approach to quantitative morphology. Bull. Geolog. Soc. Am., 56:275-370.
- Hoyt, W.G. (1942): An outline of the runoff cycle. State College, Penn., Technical Bull., 27:55-67.
- Hubbert, M.K. (1940): The theory of groundwater motion. J. Geol., 48: 785-822.
- Hursh, C.R. and Hoover, M.D. (1941): Soil profile characteristics pertinent to hydrologic studies in the southern Appalachians. Soil Sci. Soc. Am. Proc., 6:414-422.
- Hursh, C.R. (1944): Appendix B, Report of the subcommittee on subsurface flow. Eos Trans. Am. Geophys. Union, 25:743-746.

- Ivanov, A.T. (1948): Determination of the subsurface component of stream-flow by use of the hydrochemical method. Transactions of the Laboratory of Hydrogeological Problems of Acad. Sci. USSR, 3:243-246.
- Jacks, G., Olofsson, E. and Werme, G. (1986): An acid surge in a well-buffered stream. Ambio, 15(5):282-285.
- Johansson, B. (1985): A study of soil water and groundwater flow of hill-slopes - using a mathematical model. Nord. Hydrol., 16(2):67-78.
- Johansson, P.O. (1986): Diurnal groundwater level fluctuations in sandy till - A model analysis. J. Hydrol., 87:125-134.
- Johansson, S. and Nilsson, T. (1985): Hydrology of the Lake Gårdsjön area. Ecol. Bull. (Stockholm), 37:86-96.
- Jones, J.A.A. (1978): Soil pipe networks: distribution and discharge. Cambria, 5(1):1-21.
- Judy, C., Meiman, J.R. and Friedman, I. (1970): Deuterium variations in an annual snowpack. Water Resour. Res., 6:125-129.
- Kane, D.L. and Stein, J. (1983a): Field evidence of groundwater recharge in interior Alaska. In: Proc. 4th int. conf. on permafrost. National academic press, Washington D.C., 572-577.
- Kane, D.L. and Stein, J. (1983b): Water movement into seasonally frozen soils. Water Resour. Res., 19(6):1547-1557.
- Kennedy, V.C., Kendall, C., Zellweger, G.W., Wyerman, T.A. and Avanzino, R.J. (1986): Determinations of the components of stormflow using water chemistry and environmental isotopes, Mattole River basin, California. J. Hydrol., 84:107-140.
- Kirkby, M.J. (ed.) (1978): Hillslope hydrology. John Wiley & Sons, Chichester, 389 pp.
- Kirkham, D. (1947): Studies of hillside seepage in the Iowan drift area. Soil Sci. Soc. Am. Proc., 12:73-80.
- Knisel, W.G., Jr (1973): Comments on Role of subsurface flow in generating surface runoff, 2. Upstream source areas by R. Allan Freeze. Water Resour. Res., 9(4):1107-1110.
- Knutsson, G. (1971): Studies of groundwater flow in till soils. Geologiska Föreningens i Stockholm Förhandlingar, 93:1-22.
- Kobayashi, D. (1986): Separation of a snowmelt hydrograph by stream conductance. J. Hydrol., 84:157-165.
- Kramer, P.J. and Kozlowski, T.T. (1979): Physiology of woody plants. Academic Press, New York, 811 pp.
- Krouse, H.R. and Smith, J.L. (1973): $^{18}\text{O}/^{16}\text{O}$ abundance variations in Sierra Nevada. Seasonal snowpack and their use in hydrological research. In: The role of snow and ice in hydrology. Proc. Banff Symp., Sept. 1972, UNESCO-WMO-IAHS, IAHS Publ. 107:24-38.
- Liljeqvist, L. (1973): Beskrivning av Stormyra-området. SMHI, Rapport 23, 8 pp.

- Lindblad, A. (1981): Infiltrationsmätningar utförda av geologiska institutionen CTH/GU 1972-1980. Sammanställning och statistisk bearbetning. Geohydrologiska forskningsgruppen, CTH, Medd., 60, 80 pp.
- Lindström, G. and Rodhe, A. (1986): Modelling water exchange and transit times in till basins using oxygen-18. Nord. Hydrol., 17(4/5):325-334.
- Linsley, R.K. Jr, Kohler, M.A. and Paulhus, J.L.H. (1949): Applied hydrology. McGraw-Hill Book Company, Inc., New York, 689 pp.
- Lundberg, A. (1974): Infiltration och perkolation i det översta marklagret. En sammanställning av mätningar utförda i Norden. IHD Report 35, Stockholm, 58 pp.
- Lundin, L. (1982): Mark- och grundvatten i moränmark och marktypens betydelse för avrinningen. UNGI Report 56, Uppsala Univ., Dept Phys. Geogr., 216 pp.
- Luxemoore, R.J. (1983): Infiltration and runoff predictions for a grassland watershed. J. Hydrol., 65:271-278.
- Löschhorn, Von U., Ambach, W., Moser, H. and Stichler, W. (1977): Modellmässige Bestimmung von hydrologischen Verweilzeiten in einem vergletscherten Einzugsgebiet mit Hilfe von Messungen des Deuterium- und Tritiumhaltes. Z. Gletscherkd. Glazialgeol., 7(2):181-186.
- Majoube, M. (1971): Fractionnement en oxygène 18 et en deutérium entre l'eau et sa vapeur. Journal de Chimie et de Physico-Chimie Biologique, 68(19):1423-1436.
- Malmström, C. (1925): The experimental forests of Kulbäcksliden and Svartberget, 2. Vegetation. Skogsförsöksanstaltens exkursionsledare (Stockholm), 10:20-57.
- Maloszewski, P. and Zuber, A. (1982): Determining the turnover time of groundwater systems with the aid of environmental tracers, 1. Space Models and their applicability. J. Hydrol., 57:207-231.
- Maloszewski, P., Rauert, W., Stichler, W. and Herrmann, A. (1983): Application of flow models in an Alpine catchment area using tritium and deuterium data. J. Hydrol., 66:319-330.
- Martinec, J. (1974): Untersuchung der Schneeschmelze mit Umweltisotopen. Österr. Wasserwirtschaft, 26:57-61.
- Martinec, J., Siegenthaler, U., Oeschger, H. and Tongiorgi, E. (1974): New insights into the run-off mechanism by environmental isotopes. In: Isotope techniques in groundwater hydrology 1974. IAEA, Vienna, 1:129-149.
- Martinec, J. (1975): Subsurface flow from snowmelt traced by tritium. Water Resour. Res., 11:496-498.
- Martinec, J., Moser, H., De Quervain, M.R., Rauert, W. and Stichler, W. (1977): Assessment of processes in the snowpack by parallel deuterium, tritium and oxygen-18 sampling. In: Isotopes and impurities in snow and ice. Proc. Banff Symp., August-Sept. 1975, IAHS Publ. 118: 220-231.
- Megahan, W.F. (1983): Hydrologic effects of clearcutting and wildfire on steep granitic slopes in Idaho. Water Resour. Res., 19:811-819.

- Meiman, J., Friedman, I. and Hardcastle, K. (1973): Deuterium as a tracer in snow hydrology. In: The role of snow and ice in hydrology. Proc. Banff Symp., Sept. 1972, UNESCO-WHO-IAHS, IAHS Publ. 107:39-50.
- Melkerud, P.A. (1983): Quaternary deposits and bedrock outcrops in an area around Lake Gårdsjön, south-western Sweden, with physical, mineralogical and geochemical investigations. Reports in Forest Ecology and Forest Soils 44, Swedish University of Agricultural Sciences, Uppsala, 87 pp.
- Merlivat, L. and Coantic, M. (1975): Study of mass transfer at the air-water interface by an isotopic method. J. Geophys. Res., 80:3455-3464.
- Merlivat, L. (1978): Molecular diffusivities of $H_2^{16}O$, $HD^{16}O$ and $H_2^{18}O$ in gases. J. Chem. Phys., 69(6):2864-2871.
- Merot, Ph., Bourguet, M. and le Leuch, M. (1981): Analyse d'une crue a l'aide du traçage naturel par l'oxygène 18 mesurée dans les pluies, le sol, le ruisseau. Catena, 8:69-81.
- Miller, W.R. and Drever, J.I. (1977): Water chemistry of a stream following a storm, Absaroka Mountains, Wyoming. Geol. Soc. Am. Bull., 88:286-290.
- Miyake, Y. and Tsubota, H. (1963): Estimation of the direct contribution of meteoric water to river waters by means of fall-out radiocaesium and radiostrontium. Proc. Symp. Radioisotopes in Hydrology, IAEA, Vienna, 425-431.
- Mook, W.G., Groeneveld, D.J., Broun, A.E and Van Ganswijk, A.J. (1974): Analysis of a run-off hydrograph by means of natural ^{18}O . In: Isotope techniques in groundwater hydrology 1974. IAEA, Vienna, 1:145-155.
- Moser, H. and Stichler, W. (1970): Deuterium measurements on snow samples from the Alps. In: Isotope hydrology. IAEA, Vienna, 43-57.
- Moser, H. and Stichler, W. (1975): Deuterium and oxygen-18 contents as index of the properties of snow blankets. In: Snow mechanics. Proc. Grindelwald Symp., April 1974, IAHS Publ. 114:122-135.
- Moser, H. and Stichler, W. (1980): Environmental isotopes in ice and snow. In: Fritz, P. and Fontes, J-Ch. (eds), Handbook of environmental isotope geology. Vol. I, Elsevier, Amsterdam, 141-178.
- Mosley, M.P. (1979): Streamflow generation in a forested watershed, New Zealand. Water Resour. Res., 15(4):795-806.
- Mosley, M.P. (1982): Subsurface flow velocities through selected forest soils, South Island, New Zealand. J. Hydrol., 55:65-92.
- Münnich, K.O., Sonntag, C., Christmann, D. and Thoma, G. (1980): Isotope fractionation due to evaporation from sand dunes. ZfI-Mitteilungen, 29: 319-332.
- Myrabö, S. (1986): Runoff studies in a small cathment. Nord. Hydrol., 17: 335-346.
- Nakamura, R. (1971): Runoff analysis by electrical conductance of water. J. Hydrol., 14:197-212.
- Newbury, R.W., Cherry, J.A. and Cox, R.A. (1969): Groundwater-streamflow systems in Wilson Creek experimental watershed, Manitoba. Can. J. Earth Sci., 6:613-623.

- Niemczynowicz, J. (1977): Fysikalisk modell av urbaniseringspåverkat avrinningsområde. Lund Inst. Techn., Dept Water Resources Engineering, Report 3010, 53 pp.
- Nutter, W.L. (1971): The role of soil water in the hydrologic behavior of upland basins. Field soil water regime. Soil Sci. Soc. Am., special publ., 181-193.
- O'Loughlin, E.M. (1981): Saturation regions and their relations to soil and topographic properties. J. Hydrol., 53:229-246.
- O'Neil, J.R. (1968): Hydrogen and oxygen isotope fractionation between ice and water. J. Phys. Chem., 72:3683-3684.
- Olsson, B., Hallbäck, L., Johansson, S., Melkerud, P.A., Nilsson, S.I. and Nilsson, T. (1985): The Lake Gårdsjön area - physiographical and biological features. Ecol. Bull. (Stockholm), 37:10-28.
- Pearce, A.J., Stewart, M.K. and Sklash, M.G. (1986): Storm runoff generation in humid headwater catchments, 1. Where does the water come from?. Water Resour. Res., 22(8):1263-1272.
- Perrault, P. (1967): (1674). On the origin of springs. Translated by A. La Roque. Hafner Publ. Co, New York, 209 pp.
- Perttu, K., Bischof, W., Grip, H., Jansson, P-E., Lindgren, A., Lindroth, A. and Norén, B. (1980): Micrometeorology and hydrology of pine forest ecosystems, I. Field studies. In: Persson, T. (ed.), Structure and function of northern coniferous forests - an ecosystem study. Ecol. Bull. (Stockholm), 32:75-121.
- Pilgrim, D.H. (1977): Isochrones of travel time and distribution of flood storage from a tracer study on a small watershed. Water Resour. Res., 13(3):587-595.
- Pilgrim, D.H., Huff, D.D. and Steele, T.D. (1979): Use of specific conductance and contact time relations for separating flow components in storm runoff. Water Resour. Res., 15(2):329-339.
- Pinder, G.F. and Jones, J.F. (1969): Determination of the groundwater component of peak discharge from the chemistry of total runoff. Water Resour. Res., 5:438-445.
- Price, A.G., Hendrie, L.K. and Dunne, T. (1978): Controls on the production of snowmelt runoff. Proc. Modeling of Snow Cover Runoff, U.S. Army Cold Regions Research and Engineering Laboratory, Hanover, New Hampshire, Sept. 1978, 257-268.
- Price, A.G. and Hendrie, L.K (1983): Water motion in a deciduous forest during snowmelt. J. Hydrol., 64:339-356.
- Ragan, R.M. (1968): An experimental investigation of partial area contributions. IAHS Publ. 76:241-249.
- Rodhe, A. (1981): Springflood - meltwater or groundwater? Nord. Hydrol., 12(1):21-30.
- Rodhe, A. (1984): Groundwater contribution to streamflow in Swedish forested till soil as estimated by oxygen-18. In: Isotope hydrology 1983. IAEA, Vienna, 55-66.

- Rodhe, A. (1985): Groundwater contribution to streamflow in the Lake Gårdsjön area. Ecol. Bull. (Stockholm), 37:75-85.
- Rosén, K. (1982): Supply, loss and distribution of nutrients in three coniferous forest watersheds in central Sweden. Reports in Forest Ecology and Forest Soils 41, Swedish University of Agricultural Sciences, Uppsala, 70 pp.
- Rydgren, B. and Sundby, M. (1982): Ett försök att bestämma grundvattenflöde i berg med utspädningsmetoden. C-uppsats i hydrologi, Uppsala univ., 33 pp.
- Saxena, R.K. (1984): Seasonal variations of oxygen-18 in soil moisture and estimations of recharge in esker and moraine formations. Nord. Hydrol., 15:235-242.
- Saxena, R.K. (1986): Estimation of canopy reservoir capacity and oxygen-18 fractionation in throughfall in a pine forest. Nord. Hydrol., 17: 251-260.
- Saxena, R.K. (1987): Oxygen-18 fractionation in nature and estimation of groundwater recharge. Uppsala Univ., Dept Phys. Geogr., Div. Hydrol., Report Series A 40, 152 pp.
- Selby, M.J. (1973): An investigation into causes of runoff from a catchment of pumice lithology, in New Zealand. Hydrol. Sci. Bull., 18:255-280.
- Sharma, M.L., Gander, G.A. and Hunt, C.G. (1980): Spatial variability of infiltration in a watershed. J. Hydrol., 45:101-122.
- Sherman, L.K. (1932): Streamflow from rainfall by unit-graph method. Engineering news record, 108:501-505.
- Sklash, M.G. and Farvolden, R.N. (1979): The role of groundwater in storm runoff. J. Hydrol., 43:45-65.
- Sklash, M.G., Stewart, M.K. and Pearce, A.J. (1986): Storm runoff generation in humid headwater catchments, 2. A case study of hillslope and low-order stream response. Water Resour. Res., 22(8):1273-1282.
- Sokollek, V. (1985): Hydrologische Forschung im Nadelwald Nordschwedens. DFG-Bericht, Giessen, 26 pp.
- Stephenson, G.R. and Freeze, R.A. (1974): Mathematical simulation of subsurface flow contributions to snowmelt runoff, Reynolds Creek Watershed, Idaho. Water Resour. Res., 10:284-294.
- Stewart, M. (1975): Stable isotope fractionation due to evaporation and isotopic exchange of falling water drops: application to atmospheric processes and evaporation of lakes. J. Geophys. Res., 80:1138-1146.
- Stichler, W. and Herrmann, A. (1977): Variations of isotopes in snow covers as input of temperate glaciers. Z. Gletscherkd. Glazialgeol., 13: 181-191.
- Stichler, W., Rauert, W. and Martinec, J. (1981): Environmental studies of an Alpine snowpack. Nord. Hydrol., 12:297-308.
- Stichler, W. (1982): Environmental isotopes as tracers in water balance studies of mountainous watersheds. Proc. Symp. Hydrolog. Research Basins. Sonderh. Landeshydrologie, Bern, vol. II, 357-368.

- Stichler, W. and Herrmann, A. (1983): Application of environmental isotopes techniques in water balance studies of small basins. In: New approaches in water balance computations. IAHS Publ. 148:93-111.
- Tagutschi, Y. (1982): Effects of urbanization on small rivers. Beiträge zur Hydrologie, 3:203-218.
- Tamm, O. (1925): The experimental forests of Kulbäcksliden and Svartberget, 1. Geology. Skogsförsöksanstaltens exkursionsledare 10 (Stockholm), 1-19.
- Taylor, C.H. (1982): The effect on storm runoff response of seasonal variations in contributing zones in small watersheds. Nord. Hydrol., 13:165-182.
- Tennessee Valley Authority (1965): Area-stream factor correlation. Bull. Internat. Assoc. Sci. Hydrol., 10(2):22-37.
- Tóth, J. (1963): A theoretical analysis of groundwater flow in small drainage basins. J. Geophys. Res., 68:4795-4812.
- Tricker, A.S. (1981): Spatial and temporal patterns of infiltration. J. Hydrol., 49:261-277.
- Truesdell, A.H. and Hulston, J.R. (1980): Isotopic evidence on environmental of geothermal systems. In: Fritz, P. and Fontes, J-Ch. (eds). Handbook of environmental isotope geochemistry. Vol. I, Elsevier, Amsterdam, 179-226.
- Visocky, A.P. (1970): Estimating the ground-water contribution to storm runoff by the electrical conductance method. Groundwater, 8(2):5-10.
- Wakahama, G. (1968): The metamorfism of wet snow. IAHS Publ. 79:370-379.
- Walling, D.E. and Webb, B.W. (1980): The spatial dimension in the interpretation of stream solute behaviour. J. Hydrol., 47:129-149.
- Walsh, R.P.D. (1980): Runoff processes and models in humid tropics. Z. Geomorph. N.F., Suppl., 36:176-202.
- Wang, Jui H. (1965): Self-diffusion coefficients of water. J. Phys. Chem., 69:4412.
- Ward, R.C. (1984): On the response to precipitation of headwater streams in humid areas. J. Hydrol., 74:171-189.
- Wentworth, C.K. (1930): A simplified method of determining the average slope of land surfaces. Am. J. Sci., 5(20):184-194.
- Weyman, D.R. (1973): Measurements of the downslope flow of water in a soil. J. Hydrol., 20:267-288.
- Whipkey, R.Z. (1965): Subsurface stormflow from forested slopes. Bull. Internat. Assoc. Sci. Hydrol., 10(2):74-85.
- Whipkey, R.Z. and Kirkby, M.J. (1978): Flow within the soil. In: M.J. Kirkby, (ed.), Hillslope hydrology, 121-144.
- Yamagata, N., Okita, T. and Kodaira, K. (1963): The runoff of fall-out radiostrontium from a river and its application in hydrology. Proc. Symp. Radioisotopes in Hydrology, IAEA, Vienna, 417-423.

- Yurtsever, Y. and Gat, J.R. (1981): Atmospheric waters. In: Stable isotope hydrology: deuterium and oxygen in the water cycle. IAEA, Vienna, Tech. Rep. Series No. 210:103-139.
- Zekster, I.S. (1963): Role of artesian water in feeding large rivers as exemplified by the middle and lower reaches of the Neman River. Soviet Hydrol., 4:94-98.
- Zimmermann, U., Ehhalt, D. and Münnich, K-O., (1967): Soilwater movement and evapotranspiration: changes in the isotopic composition of the water. In: Isotopes in hydrology. IAEA, Vienna, 1966, 567-584.

